VOLUME XLI NUMBER 5

THE
JOURNAL OF GEOLOGY

July-August 1933

TILTING DUE TO GLACIAL MELTING

B. GUTENBERG
California Institute of Technology
Balch Graduate School of the Geological Sciences

ABSTRACT

To investigate tilting and changes in level in the United States and Canada, results from the records of tide gauges of different stations were analyzed. In the Great Lakes region all results indicate a tilting of the land upward in a northerly direction, by about 10 cm. per 100 km. per century. Along the Pacific Coast a small rising of the land is indicated at the north, whereas in California the changes in height seem to be negative, but small and irregular. Along the Atlantic Coast of Canada the changes are small and within the limits of error nearly everywhere, but south of Portland (Me.) sinking prevails clearly. It is very probable that the tilt in the Great Lakes region is due to forces which tend to restore isostatic equilibrium, disturbed by the melting of ice after the Ice Age.

In some parts of the continents which were covered by ice during the Ice Age recent uplift has been found continuing up to the present time. It is considered by many scientists, but questioned by others, that this uplift is a consequence of forces which tend to restore isostatic equilibrium disturbed by the melting of the ice. The strength of the material close to the surface of the earth is of the order of 10^9 dyn./cm.^{2}. A mountain with a height of the order of 10 km. (relative to the neighborhood) is needed to produce a vertical pressure of the order of this strength. Such an effect of glaciation therefore usually will not occur, and if tilting of the earth's crust connected with flowing movements at deeper layers is observed due to the formation of ice caps or to melting of ice, it may be concluded from this fact that the strength at depth is considerably less than

close to the surface. For this and other reasons it is very important to investigate whether any tilting observed is due to glacial melting.

The best results available on tilting of this type are around the Baltic Sea. In the early part of the eighteenth century Celsius found that the shores of this sea were rising by an average amount of 1 cm. per year. Holmström found that Södra Helsö, fronting the open Skagerrak on the west coast of Sweden, rose 30 cm. between 1820 and 1870, and Stockholm 48 cm. between 1774 and 1875. A very detailed investigation based on records of tide gauges has been

TABLE I

	North		Central Part			South		
	West	East	West	East		West	East	
	Ratan	Toppila	Väster-vik	Jung-frusund	Kron-stad	Freder-icks havn	Wismar	Memel
I	+1.2	+1.1	+0.6	+0.2	−0.7	+0.1	−0.1	+0.3
II	1.5	1.5	.7	.6	+0.1	+ .1	+ .1	+ .0
III	0.8	1.0	+ .3	.6	+1.1	− .3	− .2	+ .6
IV	+1.0	+0.6	−0.3	+0.3	+0.8	+0.5	−0.1	−0.4
Average	+1.1	+1.0	+0.4	+0.4	+0.3	+0.1	−0.1	+0.1

carried through by R. Witting.[1] He used records of more than fifty stations and corrected them for effects of meteorological changes. For the Baltic Sea he found from his data, covering fifteen years, a yearly uplift of more than 1 cm. along the northern shores, about ½ cm. in the middle part (60° N. latitude), and no change on the southern coast. The following data are taken from his tables. "I" means average yearly change of the height of the coast in centimeters, 1899–1902; "II," 1902–5; "III," 1905–8; and "IV," 1908–11. The changes were not completely regular, and Witting states that years with irregular movements show more earthquakes in these regions than occur usually.

[1] Rolf Witting, "Hafsytan," *Geoidytan och landhöjningen* (Helsingfors, 1918), Fennia XXXIX, No. 5.

The data found by Witting for points around the North Sea are very much less regular, owing probably to the fact that there the effects of storms on the height of the sea level are very much more considerable than in the Baltic Sea. The changes shown in Table II were given by Witting (periods as before, the values are centimeters per year).

TABLE II

	Aberdeen	Dundee	Vlissingen	Helder	Bremerhaven	Kattegat
I	+0.2	+0.2	+0.7	+1.9	−0.3	+0.0
II	− .3	+ .8	−2.2	−2.1	+ .3	+ .1
III	− .1	− .2	−0.8	−0.1	+ .4	− .0
IV	+0.4	+0.3	+0.8	+0.1	−0.3	+0.2
Average	0.0	+0.3	−0.4	0.0	0.0	+0.1

Witting also investigated older (but less accurate) marks of the average height of sea level in the Baltic Sea,[2] and, from observations covering a time interval of more than one hundred years, he found the following changes in height in meters per one hundred years (feet per one hundred years): 6 stations north of 62° latitude: between +0.95 and +1.00 m. (about 3 ft.); 9 stations around 60°: between +0.2 and +0.75 m. ($\frac{1}{2}$–$2\frac{1}{2}$ ft.); Swinemünde (south shore) between 0.00 and +0.05 m. (less than 2 in.).

Geological investigations show that since the end of the Ice Age the maximum rise in Scandinavia took place around the Baltic Sea between 62° and 64° latitude by an amount of over 250 m.[3] Assuming that this rise began ten thousand years ago, we find an average rise of more than $2\frac{1}{2}$ m. per century, which amount is very much larger than that which we are now observing. Witting calculated that an uplift of about 50–60 m. is still to be expected. In spite of the good agreement between calculations and observations, there is some doubt as to the extent to which these movements today are caused by the melting of the ice long ago. Especially R. Schwinner[4]

[2] "Le soulèvement récent de la Fennoscandie," *Geografisca Annaler*, 1922, p. 458.

[3] A. Born, *Isostasie und Schweremessung* (Berlin, 1923); V. Tanner, "Studier över Kvartärsystemet," *Fennoskandias mordliga delar. Bull. de la Comm. Géologique de Finlande* (Helsingfors, 1930), No. 88.

[4] "Die Schwere am Ostrand des Fennoskandischen Schildes," *Gerlands Beiträge zur Geophysik*, Vol. XXXIV (1931), p. 436.

pointed out that observations of gravity seem to contradict the idea that we have today a "postglacial uplift" in Fennoscandia.

When the ice cap formed, at depth there must have been an outflow of subcrustal material corresponding to the sinking of land in the area where the ice masses increased. That this event took place in such a way seems to be reasonable since the measurements of the thickness of ice in Greenland have shown that there is indeed a very flat trough filled with ice. One maximum height of rocks is near the coast, where in most regions heights of more than 1,000 m. occur, another seems to be in the central part. In the interior the values shown in Table III have been found.[5] When, after the Ice

TABLE III

Distance from the west coast	small	15	20	30	50	120	400 km.
Rock surface above sea level	1,000+	640	385	460	370	600–200	1,000± m.
Thickness of ice	—	330	600	750	1,200	1,600–2,000	2,000± m.

Age, the ice cap of Scandinavia had melted away, the land rose and, at some depth, the material flowed back. If this movement is still going on, then we must still have lack of material (compared with the state of isostatic equilibrium) beneath the rising region and too much material at some distance from it; this means too low gravity in the rising region, too high gravity at some distance from it. Schwinner, in his publication cited above, did not find either a well-marked region of too low gravity in the rising region, though the very scanty values there indicate a slightly negative anomaly, or a marked ring of positive anomalies around this region. Rather he found some areas with positive and other areas with negative anomalies. This could be explained by supposing that in some of these outer regions equilibrium has been restored, while in others with higher strength or viscosity there is still a remainder of the material, which flowed away from the region of glaciation when the ice caps

[5] Ernst Sorge, "Die ersten Dickenmessungen des grönländischen Inlandeises," *Zeitschrift für Geophysik*, Vol. VI (1930), p. 22; Kurt Wegener, "Die Ergebnisse der 'Deutschen Grönland Expedition Alfred Wegener,'" *Forschungen und Fortschritte*, Vol. VIII (1932), p. 143. Geophysikalische Forschungen der beiden letzten Jahre in den Polargebieten." *ibid.*, p. 422.

were formed. Besides, we must consider that the maximum anomalies which are to be expected[6] if isostatic equilibrium has not been restored since the Ice Age are probably less than 0.01 cm./sec.[2] The effects of other causes, especially the geological conditions,[7] and the errors introduced by the methods of reduction are noticeably larger than this value. Therefore, we cannot draw the conclusion that the gravity measurements disprove the theory of isostatic uplift in Scandinavia.

Conditions similar to those in Scandinavia prevailed in North America. Large parts of Canada were covered by ice during the Ice Age, which has melted away since that time. Gravity measurements are still fewer than in Europe. Small negative anomalies seem to prevail, according to A. Born,[8] in the region of former glaciation; small positive anomalies south of it. The maps showing the gravity anomalies calculated by W. Bowie[9] by use of different methods neither support nor disprove such a result, especially if we again consider the small amount of the anomalies to be expected.

Geological evidence shows that here, too, the regions which had been covered by large masses of ice during the Pleistocene Ice Age have risen since that time. The investigations especially concern the regions of the Great Lakes and around Newfoundland.[10]

In the Great Lakes region Gilbert calculated from tide-gauge rec-

[6] U. Pesonen, "Relative Bestimmungen der Schwerkraft in Finnland," *Veröff. des Finn. Geodat. Inst.* (Helsinki, 1930), No. 13, p. 162.

[7] H. Reich, "Die Bedeutung der finnischen Schweremessungen," *Gerlands Beiträge zur Geophysik, Ergänzungshefte*, Vol. II (1931), p. 1.

[8] *Op. cit.*, p. 106.

[9] "Investigations of Gravity and Isostasy," *U.S. Coast and Geodetic Surv.* (Washington, 1917), Spec. Pub. 40; "Isostatic Investigations and Data for Gravity Stations in the United States Established since 1915," *ibid.* (1924), Spec. Pub. 99.

[10] G. K. Gilbert, "Recent Earth Movement in the Great Lakes Region," *U.S. Geol. Surv., Eighteenth Ann. Rept., 1896–97*, Part II, p. 595; J. W. Goldthwait, "Isobases of the Algonquin and Iroquois Beaches," *Bull. Geol. Soc. Amer.*, Vol. XXI (1910), p. 227; F. Leverett and F. B. Taylor, "The Pleistocene of Indiana and Michigan and the History of the Great Lakes," *U.S. Geol. Surv., Monog. 53* (1915); J. W. Spencer, "Postglacial Earth Movements about Lake Ontario . . . .," *Bull. Geol. Soc. Amer.*, Vol. XXIV (1913), p. 217; R. A. Daly, "Oscillations of Level . . . .," *Bull. Geol. Soc. Amer.*, Vol. XXXI (1920), p. 308; "Post-glacial Warping of Newfoundland and Nova Scotia," *Amer. Jour. Sci.* (5th ser.), Vol. I (1921), p. 381; "Pleistocene Changes of Level," *ibid.*, Vol. X (1925), p. 281.

ords the changes shown in Table IV in height difference. From these data he derived an average change in slope of 0.42±0.044 feet (12 cm.) per 100 miles per century. This maximum tilting is supposed to be upward from S.SW. to N.NE. Spencer, on the other hand, considered that these results are erroneous and that there is no tilting now going on in the lake region. He calculated the mean water level

TABLE IV

Sacketts Harbor–Charlotte.....	1874–96	−0.061 ft.	(±0.03)
Cleveland–Pt. Colborne.....	1858–95	− .239	(.06)
Milwaukee–Pt. Austin.....	1876–96	− .138	(.03)
Escanaba-Milwaukee.....	1876–96	+0.173	(±0.02)

for quinquennial periods between 1855–59 and 1906–10 at Colborne and at Cleveland and found by this less accurate method no noticeable change during this whole time interval.

Proof that these movements are still going on has been given by Sherman Moore[11] and John R. Freeman.[12] Moore used the readings of nineteen pairs of gauges during the five months June to October as most free of disturbances.

> Differences in the elevation of the water surface for each season between all possible pairs of gauges on each lake were then taken out, plotted on cross-section paper, and a line drawn through the points. . . . . The slope of this line determined the probable rate of change. . . . . Equations were then written of the form
>
> $$ax+by-c=0\,,$$
>
> in which a and b were the distances in miles north and east respectively between the gauges, c was the observed rate of change in elevation . . . . and x and y were unknown quantities, the north and east components of the rate of movement.

He found the results shown in Table V.

In conclusion Moore stated that the present axis of tilt is approximately W. 20° N. (maximum uplift toward N. 20° E.), still in the same general direction as it has been since the formation of the old beaches, and that the rate is not uniform, but is about twice as

[11] "Tilt of the Earth in the Great Lakes Region," *Military Engineer*, Vol. XIV (May-June, 1922), p. 151.

[12] *Regulation of the Great Lakes* (Chicago Sanitary District, 1926), pp. 149–72.

great in the Superior Basin and three times as great in the Ontario Basin as in the remainder of the region.

Freeman used twenty pairs of gauge comparisons. By a thorough examination he found that "continuous progressive tilting upward toward the north at the rate of about half a foot per 100 miles per century in the southern part of the Lake region, with indications of double this rate over some parts of the Lake system, is proved beyond all doubt." His results are given in Figure 1.

As results of gauge readings have accumulated since the investigation of Freeman, especially concerning Lake Superior, where Moore

TABLE V

Lake	Number of Equations	Maximum Tilt per Century		
		Amount Ft / 100 Mi.	Cm./100 km.	Direction To
All four lakes	19	0.43	8.1	N. 25° E.
Superior	3	(0.94)	(17.8)	Indefinite
Michigan-Huron	9	0.47	8.9	N. 18 E.
Erie	3	0.46	8.7	N. 31 E.
Ontario	4	1.44	27.3	N. 18 E.

and Freeman could use only three stations in one line, so that the direction of tilt was indefinite, it has seemed worth while to recalculate the changes in level by using all data available. Gauge readings were provided through the courtesy of the United States Lake Survey Office at Detroit, the Canadian Hydrographic Service of the Department of Marine at Ottawa, and the United States Engineer Office at Duluth. For each lake certain pairs of stations were used, especially those covering long periods and running in different directions. It is assumed that the difference in the gauge readings of each two stations are given by $a+bt$, where t is the time, a the difference in the gauge readings at the time $t=0$, which usually was taken in the middle of the interval, and b the average change in this reading per unit of t (year). a and b were calculated by means of the method of least squares for all pairs of stations given in Table VI, except the two pairs concerning Escanaba, which were taken from the calcula-

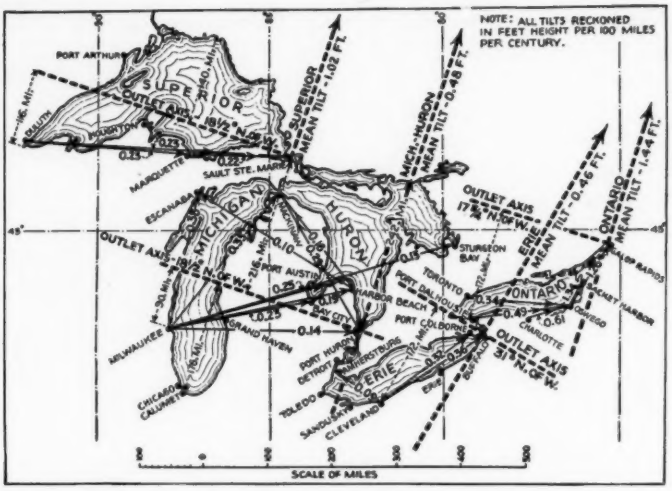

FIG. 1.—Rates and directions of earth tilt for each of the Great Lakes. (Taken from John R. Freeman, *op. cit.*, p. 150.)

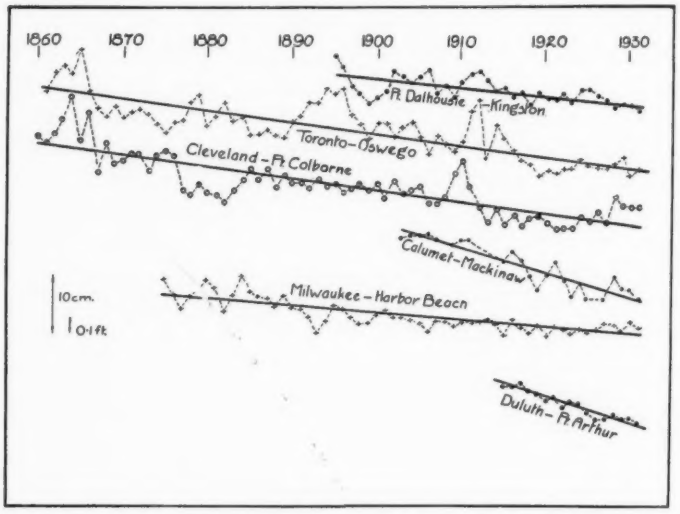

FIG. 2.—Changes in differences in gauge readings in the Great Lakes (cf. Table III and Fig. 3).

tions of Freeman, and the pair Sault Ste Marie–Marquette, taken from the results of Moore.

The changes given in Table VI were plotted on a map (Fig. 3). It is not very difficult to draw lines of equal changes in level, as has

TABLE VI*

CHANGES IN DIFFERENCES OF GAUGE READINGS

LAKE	STATIONS	CHANGE PER CENTURY		PERIOD USED FOR CALCULATION
		Ft.	Cm.	
Superior	Duluth–Pt. Arthur	1.49±0.12	45± 4	1915–1931
Superior	Duluth-Marquette	0.59±0.06	18± 2	1889–1931
Superior	Marquette–Pt. Arthur	0.97±0.11	30± 3	1915–1931
Superior	Marquette-Michipicoten	0.70±0.23	21± 7	1918–1931
Superior	Michipicoten–Pt. Arthur	0.25±0.25	8± 8	1918–1931
Superior	Marquette–Sault Ste Marie	0.39± ?	12± ?	1872–1919
Michigan-Huron	Milwaukee-Escanaba	0.75± ?	23± ?	?
Michigan-Huron	Milwaukee-Mackinaw	0.60±0.07	18± 2	1900–1931
Michigan-Huron	Calumet-Milwaukee	0.98±0.21	30± 4	1903–1931
Michigan-Huron	Calumet-Mackinaw	1.46±0.14	44± 4	1903–1931
Michigan-Huron	Calumet–Harbor Beach	1.02±0.14	31± 4	1903–1931
Michigan-Huron	Milwaukee–Harbor Beach	0.41±0.03	12± 1	1875–1931
Michigan-Huron	Harbor Beach–Escanaba	0.27± ?	9± ?	?
Michigan-Huron	Mackinaw-Collingwood	0.05±0.19	2± 6	1916–1931
Michigan-Huron	Harbor Beach–Collingwood	0.55±0.13	17± 4	1916–1931
Michigan-Huron	Calumet-Collingwood	1.30±0.38	40±12	1916–1931
Michigan-Huron	Harbor Beach–Goderich	0.13±0.19	4± 6	1920–1931
Michigan-Huron	Goderich-Collingwood	0.47±0.11	17± 4	1920–1931
Michigan-Huron	Harbor Beach–Mackinaw	0.27±0.05	8± 2	1900–1931
Erie	Cleveland-Amherstburg	0.07±0.15	2± 5	1899–1918
Erie	Cleveland–Pt. Colborne	0.67±0.05	20± 2	1860–1931
Erie	Cleveland-Buffalo	0.62±0.06	19± 2	1887–1931
Ontario	Toronto–Pt. Dalhousie	0.16±0.08	5± 2	1861–1931
Ontario	Toronto-Oswego	0.64±0.06	19± 2	1861–1931
Ontario	Toronto–Cape Vincent	1.35±0.19	41± 6	1900–1931
Ontario	Pt. Dalhousie–Kingston	0.43±0.09	13± 3	1895–1931
Ontario	Kingston–Cape Vincent	0.65±0.07	20± 2	1900–1931
Ontario	Oswego–Cape Vincent	0.48±0.09	15± 3	1900–1931

* ± indicates the mean error. Decrease of water level (uplift of land) in the direction toward the station at the second place. Some special data are shown in Fig. 2.

been done in Figure 3. The position of the zero line is arbitrary; it is drawn through Calumet, where the least upheaval (or the largest subsidence) occurs of all points used in Figure 3. The average tilt per century is about ½ foot per 100 miles, or about 10 cm. per 100 km. The same values are found in Scandinavia.

The fact that different results do not check completely is due to the use of data covering different periods. In some cases they may

be influenced by instability of bench marks, but the fact that stations close together and stations situated on lines nearly parallel to the axis of uplift show no great changes in the difference of level shows that the errors caused in this way usually are very small. The large tilt between Calumet and Milwaukee has looked suspicious to all investigators, but it has been determined on several occasions that the gauge at Calumet Harbor has been stable. Another unusual feature is the considerable uplift at Cape Vincent. The self-registering

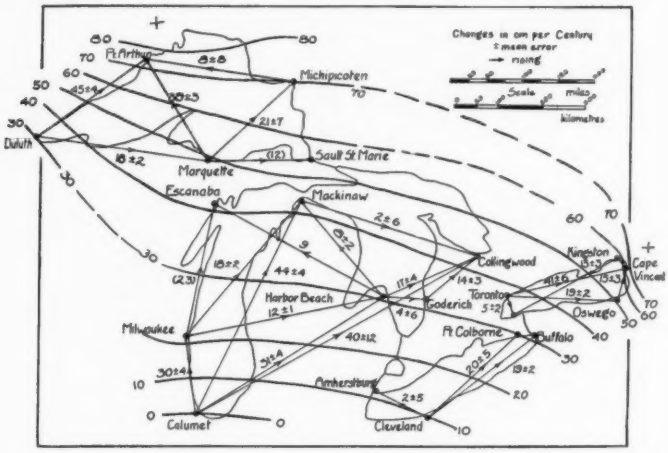

FIG. 3.—Actual changes in level in the Great Lakes region in centimeters per century. (Arrows point in direction of uplift; lines of equal uplift from 10 to 10 cm.)

gauge was removed in 1915 from Tibetts Point to Cape Vincent, but this fact is not indicated by the results. The three pairs of sets of observations concerning this station run very smoothly and the final errors are below average. The calculated uplift of Cape Vincent compared with Kingston, for example, is 0.20 feet between 1900 and 1931, the calculated errors for three years are −0.12, +0.11, and +0.06 feet, while in all other years their absolute amount is less than 0.06 feet. The statement by Freeman that this method of finding changes in level is more accurate than geodetic leveling is true without any doubt.

Another possibility of finding changes in level is through examination of the readings of tide gauges at the coasts, though these are very much more affected by the tides and meteorological conditions than the readings of gauges on the shores of lakes.

In the case of Newfoundland, geological evidence shows an uplift in recent time. On the other hand, W. B. Dawson found[13] from tide-gauge records in 1917 that at that time there was no measurable change in height of the land at Halifax, Charlottetown, and St. Paul Island (Cabot Strait), and R. A. Daly[14] drew the conclusion that the New England Coast seems to have been sensibly stable for at least one hundred years. More recent data on the Tide Levels and datum planes for Eastern Canada and the Pacific Canadian Coast were provided by courtesy of the Canadian Hydrographic Service (Tidal Division) of the Department of Marine, Ottawa. Mr. F. Anderson, hydrographer of this Department, stated that the only instance in which there has been a direct evidence from tidal observations of an upheaval occurred in the lower St. Lawrence region during an earthquake.

Similar values for the United States were provided by courtesy of the United States Coast and Geodetic Survey.. The data concerning Canada are based on mean yearly values, calculated by the Hydrographic Service of Canada (Marine Dept.) from the harmonic analysis of the hourly ordinates of the records of tide gauges, whereas the data concerning the United States have been calculated by the United States Coast and Geodetic Survey by averaging the hourly heights for that year; they have not been subjected to harmonic analysis.

In Table VII the changes between the first complete period of five years, or if incomplete at least three years, and the last period of five years are given as derived from these data just mentioned. To get more dependable results from all data at each station, the change in sea level has been calculated by the method of least squares and under the assumption that the mean sea level undergoes changes proportional to the time (Fig. 4). These results are given

[13] *Tide Levels and Datum Planes in Eastern Canada* (Ottawa: Department of the Naval Service, 1917).

[14] "Post-glacial Warping . . . . ," *op. cit.*, p. 390.

in Table VIII. If the mean error, indicated by ± in the table, is equal to the calculated change, the probability that the direction of the change is true is about 92 per cent, supposing that there is no systematic error; if the calculated change is twice the mean error, the corresponding probability is 98½ per cent, and in case of a result equal to three times the mean error, it is 99.9 per cent. Therefore, calculated changes less than the mean error, at one station only, do

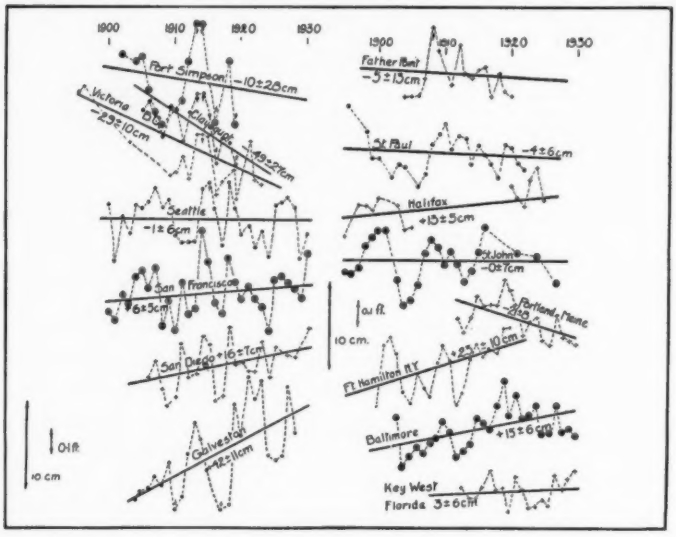

FIG. 4.—Changes in sea level

not indicate any decisive result; if the amplitude of the changes is of the order of twice the mean error, the direction is very probable, especially if similar values occur in the neighborhood, and results larger than three times the mean error indicate that a change of that order in the calculated direction is nearly beyond doubt, supposing that there are no systematic errors, as, for example, effects of very long-period tides or long-period meteorologic changes. The amplitudes of long-period tides are usually very much less than the errors calculated in Table VIII, and according to data collected by Freeman very probably this holds too in the case of effects of changes in

pressure over the Atlantic Ocean. Considering all these facts, we see that there has been no noticeable change in sea level along the eastern coast of Canada. Along the southeastern coast of the United States there is a clear though not very rapid rise in sea level, corresponding to a sinking of the coast. The maximum calculated in Table VIII occurs at Galveston; its amount is about $\frac{1}{2}$ cm. per year,

TABLE VII

CHANGES IN SEA LEVEL IN FEET
(1 ft. = 30.5 cm.; 1 cm. = *ca.* 0.033 ft.)

Atlantic Coast		
Father Pt.	1904/1909–1916/20	−0.05
St. Paul	1898/1900–1919/22	− .03
Halifax	1897/1900–1920/25	+ .06
St. John	1895/1901–1913/16	−0.02
Portland, Me.	1912/16 –1926/30	−0.07
Ft. Hamilton, N.Y.	1900/05 –1916/20	+ .11
Baltimore, Md.	1903/07 –1926/30	+ .08
Key West, Fla.	1913/17 –1926/30	+ .02
Galveston, Tex.	1904/08 –1924/28	+0.19
Pacific Coast		
Pt. Simpson	1904/08 –1916/18, 19	−0.09
Clayoquot	1908/10 –1915/19	− .17
Vancouver	1905/09 –1921/23	+ .03
Victoria, B.C.	1909/13 –1919/22	−0.04
Seattle, Wash.	1900/04 –1926/30	+0.03
San Francisco	1900/04 –1926/30	.08
San Diego, Calif.	1906/10 –1927, 28, 30	+0.15

or about $\frac{1}{5}$ inch per year. More data are needed to get more accurate results, and it would be very useful for all investigations of this kind, which are important in many respects, if more tide gauges could be installed along all coasts.

Results relative to the northern part of the Pacific Coast show a decrease in sea level. The individual figures are not decisive, as the errors are rather large. This is caused partly by the small number of data (e.g., thirteen yearly averages from Port Simpson and fourteen from Clayoquot) and partly by the fact that there is a region of very

large change in air pressure caused by the low-pressure areas which frequently pass over this region. Therefore, on account of meteorological effects, the yearly averages of sea level differ very much more than they do anywhere else in the region considered in Table VIII.

TABLE VIII

CHANGES IN SEA LEVEL CALCULATED BY MEANS OF THE METHOD OF LEAST SQUARES

Location of Tide Gauge	Period of Observ.	Change Ft./100 Yrs.	Cm./100 Yrs.
Atlantic Coast (from North to South)			
Father Pt. (St. Lawrence estuary)	1904–1920†	−0.2±0.4	− 5±13
St. Paul (Cabot Strait)	1895–1922†	−0.1±0.2	− 4± 6
Halifax	1895–1925†	+0.4±0.2	+13± 5
St. John (Bay of Fundy, lower part)	1895–1927†	−0.0±0.2	− 1± 7
Portland, Me.	1912–1930	−0.7±0.3	−21± 8
Ft. Hamilton, N.Y.	1900–1920	+0.8±0.3	+23±10
Baltimore, Md.	1903–1930	+0.5±0.2	+15± 6
Key West, Fla.	1913–1930	+0.1±0.3	+ 3± 6
Galveston, Tex.	1904–1928	+1.8±0.4	+54±13
Pacific Coast (from North to South)			
Pt. Simpson (northern B.C.)	1902–1919†	−0.3±0.9	−10±28
Clayoquot,* B.C.	1905–1919†	−1.6±0.9	−49±27
Vancouver, B.C.	1903–1923†	−0.1±0.4	− 2±13
Victoria, B.C.	1895–1922†	−0.9±0.3	−29±10
Seattle, Wash.	1900–1930	−0.0±0.2	− 1± 6
San Francisco, Calif.	1900–1930	+0.2±0.2	+ 6± 5
San Diego, Calif.	1906–1930†	+0.5±0.2	+16± 7

* West coast of Vancouver Island.
† Incomplete.

For example, very large differences occurred during the period 1913–16 at the stations on the North Pacific Coast. The calculated "errors" are (in 1/100 ft.) as indicated in Table IX. These "errors" change gradually from station to station, and there can be no doubt that they are influenced by meteorological effects. This has already been found by W. B. Dawson.[15]

Owing to this fact, a part of the "errors" will be eliminated if the

[15] *Tide Levels and Datum Planes on the Pacific Coast of Canada* (Ottawa: Department of Marine and Fisheries, 1923).

differences between the levels at the different stations are used in a similar way as has been done in the case of the data concerning the Great Lakes. But unfortunately the data cover different periods, so that this method can be used in a few cases only, and even in these the number of values is considerably less than in the case when each

TABLE IX

Year	Pt. Simps.	Clay.	Vanc.	Vict.	Seat.	San Franc.	San Diego
1912	+ 8	+ 4	− 2	−10	− 9	− 7	− 2
1913	+27	+20	+ 5	+ 7	− 9	− 6	− 5
1914	+28	+29	+15	+21	+11	+24	+13
1915	?	+ 8	+10	+ 9	+18	+12	+ 9
1916	−23	−24	−20	−17	+ 4	− 5	−12
1917	?	+ 2	+ 1	− 7	−16	− 9	−11
1918	+11	+ 2	+12	+20	+14	+12	+14

TABLE X

Stations	Change in Difference in Sea Level*	
	Ft.	Cm/100 Yrs.
Clayoquot-Vancouver	−1.4±0.6	−42±17
Vancouver-Seattle	+0.1±0.7	+ 4±21
Seattle–San Francisco	−0.2±0.3	− 7± 8
San Francisco–San Diego	−0.3±0.2	− 9± 6

* + means that the water at the first is rising, or that the land there is sinking compared with the second.

station is considered separately. The results shown in Table X have been found in this way. The results confirm those of Table VIII.

The rise of land along the northern part of this coast seems to be a very few millimeters per year (about $\frac{1}{10}$ in.). At the two Californian stations the figures indicate a very small sinking of the land during the last twenty-five years, not very much exceeding the mean error of the results.

It is very important that investigations of other kinds and covering earlier periods have been made in a few cases. There are, es-

pecially, two investigations by John R. Freeman,[16] the first concerning the vicinity of New York and the second concerning Boston. Using all the data on height of mean sea level around New York from 1843 to 1902, Freeman, Burr, and Hering found that during the sixty years from 1843 to 1902 the apparent subsidence was 0.45 foot, or at the rate of 9 inches (23 cm.) in one hundred years. It is a coincidence that we find exactly the same figure in Table VIII from the records 1900–1920, but this makes it nearly certain that there was a noticeable subsidence of land during the last one hundred years. According to the second very detailed study by Freeman, the land around Boston is sinking at the rate of 1 foot (30 cm.) in one hundred years. In 1930 the datum plane to which all elevations are referred by the engineering department ("Boston base") probably coincided almost exactly with mean low water at the Charleston navy yard; in 1878 the Boston base was found to be 0.64 foot below mean water, and in 1902 0.79 foot below. Making a similar comparison on the basis of mean sea level, Freeman finds a change of 0.71 foot in seventy-two years. Besides, he quotes similar results from many other observations (e.g., water level near rocks, soundings) around Boston, found by W. O. Crosby and John H. Sears. The point nearest to Boston, from which we have data in Table VIII, is Portland (Me.), about 160 km, to the north, and there we find a probable uplift of the land.

Along the Pacific Coast there is geological evidence for recent gradual changes, as has been found, for example, by Buwalda[17] and Wood.[18] But in this region traversed by many fault zones apparently the amount and even the direction of the movements change very much more with distance and time than they do in the Atlantic region. Gradual (and abrupt) block movements seem to prevail.

[16] W. H. Burr, R. Hering, and J. F. Freeman, *Report of the Commission on Additional Water Supply for the City of New York* (New York, 1904), p. 650; John R. Freeman, "Subsidence of Land and Harbour Bottom," Appendix 20 to the *Report of the Committee on Charles River Dam* (Boston: State of Massachusetts, 1903), pp. 529–72.

[17] John P. Buwalda, "Nature of the Late Movements on the Haywards Rift," *Central Calif. Bull. Seism. Soc. Amer.*, Vol. XIX (1929), p. 187.

[18] H. O. Wood, "On a Possible Causal Mechanism for Heave-Fault Slipping in the California Coast Range Region," *Bull. Seism. Soc. Amer.*, Vol. V (1915), pp. 214–29.

Geodetic measurements have been carried out and repeated in a few cases, but the time elapsed since these investigations have been started is not long enough to give decisive results. A comparison of triangulation made before 1900 and repeated between 1922 and 1925 has been calculated by Bowie.[19] This shows that during this time interval noticeable horizontal movements occurred only in the region adjacent to the San Andreas fault and very probably not exceeding 7 feet (2 m.); how far these movements originated during the earthquake of San Francisco in 1906 is not known.

From an investigation on the levels in Los Angeles Harbor region G. F. Nicholson[20] found an uplift of land in the northern part of the harbor of about 0.2 foot (6 cm.) between 1919 and 1927, under the supposition that a bench mark on Deadman's Island in the southern part did not change. It is not impossible that the limit between the area of small and that of noticeable change is determined by a fault, as Nicholson believes.

If we try to sum up the results, we see that along the Pacific Coast of North America, especially in California, gradual changes seem to be very much more irregular than along the Atlantic Coast. In Western Canada perhaps a small rise of land may be going on. The eastern part of North America seems to show a rise of land in the north caused probably by uprising following the melting of ice after the Ice Age, which is very clearly marked in the Great Lakes region; also by geological evidence, rather than by recent measurements, along the Atlantic Coast, where recent movements seem to be less than ± 5 cm. during the last twenty-five years. Along the southeastern coast of the United States there seem to be subsidences of land by different amounts—for example, about $\frac{3}{4}$ foot ($\frac{1}{4}$ m.) in one hundred years near New York and perhaps twice as much near Galveston (Tex.) but nowhere has any indication been found there of vertical movements larger than $\frac{1}{2}$ cm. per year.

It is of interest that the changes in level found for the west coast

[19] "Comparison of Old and New Triangulation in California," *op. cit.* (Washington, 1928), Spec. Pub. 151.

[20] "Variation in Levels, 1919 to 1927, in Los Angeles Harbor," *Bull. Seis. Soc. Amer.*, Vol. XIX (1929), p. 200.

of Europe are very similar to those found for the east coast of North America. Figure 5[21] shows in Northern Europe the uplift of Scandinavia which we considered above. The data concerning France are

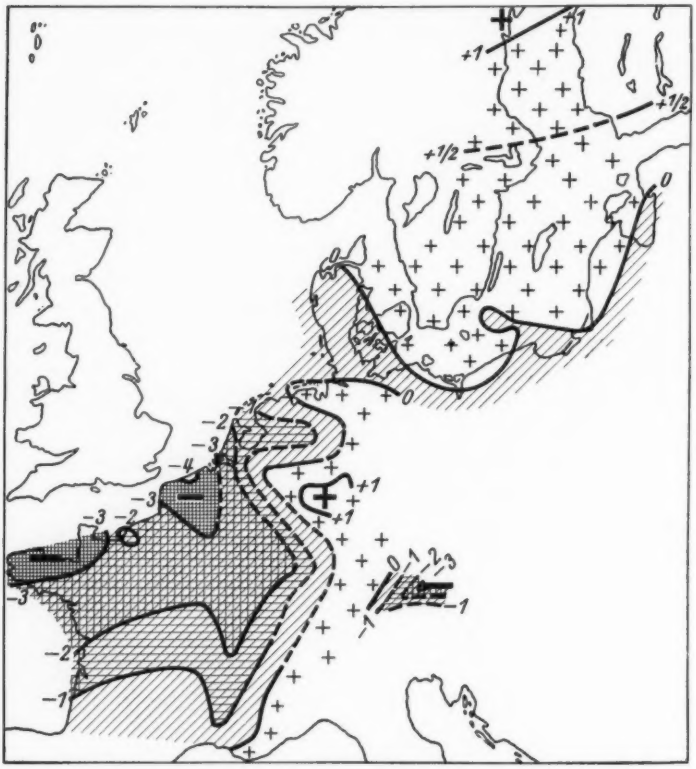

FIG. 5.—Vertical movements in Western Europe (cm. ±/year)

based on a comparison of levelings between 1857–64 and 1884–93, calculated by M. Schmidt.[22] There is some doubt as to what extent

[21] From Gutenberg, "Geotektonische Hypothesen," *Handbuch der Geophysik* (Berlin: Verlag Gebr. Borntraeger, 1930), Vol. III, p. 462.

[22] "Neuzeitliche Erdkrustenbewegungen in Frankreich," *Sitzungsberichte der Bayrischen Akademie der Wissenschaften* (München: Math.-phys. Klasse, 1922), p. 1. The connection of these changes with the geology has been discussed by Em. Kayser in the same volume (p. 51).

these results are reliable. The maximum subsidence of land would correspond to nearly 15 feet per one hundred years. On the other hand, the changes are so large that their direction at least seems to be true, and, in addition, they do not disagree with the data which have been found for other parts of Central Europe,[23] and which have been used in drawing Figure 5. In Bavaria, Schmidt (*loc. cit.*) found, besides, horizontal movements from triangulation amounting to about 4 m. (13 ft.) in one hundred years at maximum.

It is, of course, very difficult to explain these changes. The uplift of Scandinavia and of parts of Northern North America is caused probably by forces still tending to restore isostatic equilibrium, which has been disturbed by the melting of ice after the Ice Age. Gravity measurements do not contradict this hypothesis, as we have seen. The results of geological investigations which seem to show that the uplift since the Ice Age is about proportional to the probable thickness of the ice at that time support this idea considerably. The general sinking of the coasts of the northern Atlantic Ocean may be due to the process which formed the Atlantic Ocean and perhaps is still going on, but no emphasis is put upon this explanation.[24] On the other hand, it does not seem to be very probable that the sinking of the coasts in France and the Southeastern United States is due to the fact that material is still flowing at depth from beneath these regions under the rising regions in the north, as the maximum uplift seems to be too far to the south in both cases, and the volume of the outflowing mass would be very much larger than the corresponding volume of the masses in the rising regions.

[23] J. L. Wilser, *Heutige Bewegungen der Erdkruste, erkennbar an Ingenieurbauten im Oberrhein-Gebiet*, (Stuttgart, 1929); M. Schmidt, "Untersuchungen von Höhen- und Lagenänderungen im bayrischen Alpenvorland" (*op. cit.* München: Akademie, 1918), p. 373; H. Schütte, "Krustenbewegungen an der deutschen Nordseeküste," *Aus der Heimat*, Vol. XL (1927), p. 325; J. Weissner, *Der Nachweis jüngster tektonischer Bodenbewegungen im Rheinland* (Haarfelddruck, 1929); Gutenberg, *op. cit.*, p. 458.

[24] B. Gutenberg, "Hypotheses on the Development of the Earth," *Jour. Wash. Acad. Sci.*, Vol. XX (1930), p. 21.

PETROLOGY OF THE CAMBRIAN-ORDOVICIAN CONTACT IN MINNESOTA[1]

WILLIAM A. P. GRAHAM
Ohio State University

ABSTRACT

The Jordan sandstone–Oneota dolomite contact was studied in detail in the field, and numerous samples were collected for microscopic study. Evidence in support of solution of the dolomite at the contact and of the presence of a large diastem between these two formations is presented.

INTRODUCTION

For several years the contact relationship in Minnesota and Wisconsin between the Cambrian and Ordovician formations—Jordan sandstone and Oneota dolomite, respectively—has been considered an unconformable one by some geologists,[2] other investigators consider these two formations conformable; while still others maintain that a diastem separates them.

Certain features occur in this region which at first sight are not compatible with a single interpretation of the observed phenomena. These phenomena are listed below under the relationship they most strongly support. (I) Conformability is suggested by: (1) the apparent gradation of the sandstone into the dolomite in the southern part of the area; (2) absence of recognizable relief at the top of the Jordan sandstone; (3) general decrease in coarseness of grain size as the dolomite is approached. (II) A break with the magnitude of a large diastem is suggested by: (1) the absence of a notable change in the mineralogy of the heavy concentrates; (2) the brecciation of the friable sandstone some distance below the dolomite; (3) the presence of brecciation only in the northern half of the area; (4) the sporadic occurrence of iron oxide and clay between the sandstone and dolomite, interpreted as being a result of solution of the basal Oneota in

[1] The writer wishes to express appreciation to Dr. C. R. Stauffer and Dr. A. C. Trowbridge for helpful discussion on several points during the preparation of this paper.

[2] E. O. Ulrich, "Notes on New Names in Table of Formations and on Physical Evidence of Breaks between Paleozoic Systems in Wisconsin," *Trans. Wis. Acad. of Sci., Arts, and Letters*, Vol. XXI (1924), p. 105; C. R. Stauffer, "The Jordan Sandstone," *Jour. Geol.*, Vol. XXXIII (1925), p. 706.

post-Oneota time rather than a residual subaërial accumulation; (5) a gradual increase in the amount of dolomitic cement in the sandstones above the brecciated zone as the massive dolomite is approached. (That this cement was not always introduced secondarily by solution of the dolomite is shown by the occurrence of friable sandstone beds between well-cemented sandstone beds.) (III) An unconformity is suggested by the following: (1) the presence of ferruginous clay in some outcrops suggesting subaërial accumulation on an old land surface;[3] (2) the thick Ozarkian deposits of Missouri and elsewhere suggest missing members in this area; (3) the very sharp contact at some outcrops might well be explained by a major break in sedimentation; (4) discovery of fossils in the sandstones above the brecciated zone and below the dolomite, which are at present interpreted as being neither of Jordan nor Oneota age.[4]

This paper presents the results of a petrographic study of the contact phases of these two formations, as well as new field data, in an attempt to show that, although there is in places a definite break in sedimentation at the contact, the amount of erosion was such as to indicate a relatively short erosion interval.

The area covered in this investigation is shown in Figure 1. The location of the sections is indicated by the various locality names. The locality numbers used on the chart designate the geographic location of each series of samples.

FIELD RELATIONS

At some localities the contact between the Jordan sandstone and Oneota dolomite is extremely sharp; whereas at others a gradation from the sandstone into sandy dolomite occurs. The sharp contacts are most frequently found in the northern portion of the area, e.g., at Stillwater, Ottawa, and Mankato; whereas gradational contacts occur mostly in the more southern portions of the state. Where the contact is sharp, it is usually wavy, the individual undulations having an amplitude and wave length varying from a few inches to a few feet.

[3] E. O. Ulrich, "Revision of the Paleozoic Systems," *Bull. Geol. Soc. Amer.*, Vol. XXII (1911), pp. 459–61.

[4] C. R. Stauffer, personal communication.

At two widely separated locations—Kasota and Red Wing—a brecciated zone at the upper Jordan contact occurs from 20 to 30 feet below the first definite dolomite horizon of the Oneota. At the Kasota locality post-Jordan fossils have been found in the sandstone immediately above the brecciated zone,[5] but at Red Wing no fossils were found. On the basis of field evidence in these localities it seems imperative that the contact be placed at the base of the brecciated zone, since the evidence is clear that the advancing Oneota sea reworked the upper portion of the Jordan sandstone.

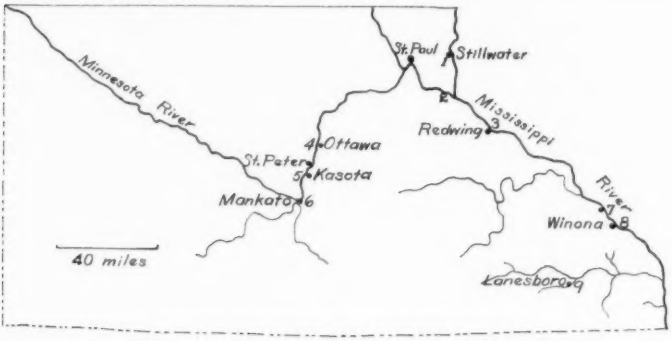

FIG. 1.—The area investigated with the location of the sections studied in detail noted by numbers. The numbers on the map correspond to the section numbers in Table I.

Powell,[6] in studying the Oneota fauna, discovered at St. Peter and Kasota a few fossils which he considers typical forms of neither the Jordan nor Oneota. At Kasota these forms occur in the sandstone above the breccia, and at St. Peter in the sandy zone immediately below the dolomite. The following four possible explanations might be offered to account for this rather unusual find: (1) Since the fauna is not well preserved, there may be an error in identification. (2) These forms may represent a very localized divergent evolutionary tendency of early Oneota life. The two localities supplying the entire new fauna are but a very few miles apart. (3) A change in lithology of a deposit is normally accompanied by a change in fauna. At the two localities in question a lithological change occurs in passing from

[5] Stauffer, *op. cit.*, p. 706.

[6] Louis H. Powell, manuscript in preparation.

a sandstone to a dolomite. (4) The fauna may represent an earlier Ozarkian formation than the Oneota, not previously recognized. The physical appearance of the zone in which these new forms occur resembles closely the Madison sandstone of Wisconsin and occurs in the proper stratigraphic position to be so correlated, if it is eventually shown to be distinct from the Oneota. It is not believed, at the present time, that these new fossils have been sufficiently well established to warrant considering the unit in which they occur anything other than basal Oneota.

In the neighborhood of Mankato the Oneota dolomite is well jointed wherever exposed and shows evidence of solution along these joints near and at the contact with the Jordan. The effect of solution along the joints is shown by the smooth, rounded, and somewhat undulatory joint surfaces, and also by the concentration of iron oxide so frequently found coating the surface of the dolomite in these channelways. On reaching the sandstone the waters spread laterally, developing small solution cavities at the base of the dolomite, over rather extended areas. This phenomenon would be especially developed at times when the water table closely approached the contact plane between the two formations.

The top of the Jordan, shown in Figure 2, is a fine-grained, water-bearing sandstone, the fineness of the texture causing the water to be held for some time after the coarser underlying layers are dry. Water passing downward along the joints in the Oneota is slowed up in its downward motion on reaching the fine sandstone, resulting in lateral spreading and giving rise to the conditions necessary for the accomplishment of the observed solution effects. At St. Peter a medium-textured phase of the Jordan is in contact with the Oneota. The sandstone is massive, thus masking the irregularities in the bedding planes, although the irregular re-entrants at the contact are very well developed. Downward-moving waters would not be slowed up under such conditions on reaching the sandstone unless the water table roughly coincided with the contact of the sandstone and dolomite. Since extensive solution did occur under this condition, the foregoing relationship is assumed to have existed at some time prior to the settling of the dolomite into its present position.

As solution progressed, the sandstone below was deformed and

squeezed into its present position by settling of the dolomite due to irregular sinking of the unsupported and weakened portions of the dolomite. In some cases the sandstone has been forced 15–20 feet upward along joints in the dolomite; in places it no doubt was forced even greater distance upward, but erosion has removed the beds so no greater measurements were secured. It is assumed that settling of the dolomite approximately kept pace with solution preventing the accumulation of irregular dolomite blocks on the floor of the caverns, since no such blocks have been found. The

FIG. 2.—Squeeze at the contact of the Oneota dolomite and Jordan sandstone, resulting from solution of the dolomite with accompanying settling of the dolomite into the sandstone. Sibley Park, Mankato, Minnesota.

deformation of the very fine-grained friable sandstone or silt zone at the top of the Jordan is expressed by an irregular series of small folds definitely showing thinning on the flanks with thickening along the axes. These folds have their axes under joints in the dolomite which have been filled by the injection of sandstone during the period of settling of the blocks. The Oneota is invariably broken and disturbed where these clay and sandstone re-entrants and joint fillings occur, the intensity and size of the fractures in the dolomite being proportional to the squeezing displayed in the sandstone.

Winchell and Upham[7] noted these irregularities at this contact

[7] Warren Upham, *Geol. and Nat. Hist. Surv. of Minn.*, Vol. I (1872–82), pp. 432–38.

and suggested downward migration of Cretaceous clays along joints in the Oneota during Cretaceous time. They assumed the clays were then carried into large and widespread solution cavities at the base of the Oneota, where the stratification developed conformed to the shape of the openings. Where these clays, or sandy silts, have been observed below the Oneota in the present study, their stratification is parallel to that in the Jordan, except at the localities where they show the result of squeezing and injection upward into the joints of the Oneota. The stratification in the sandy silts always bends downward at the contact with the dolomite instead of upward, as would be the case if the material were introduced from above.

Just when the Oneota settled into the Upper Jordan cannot be definitely stated, but it was no doubt when the competency of the dolomite was low, owing either to a sufficient reduction in thickness of the poorly supported formation or to the added weight of one of the Pleistocene ice sheets. It is thought that settling took place in post-Tertiary time, but no evidence can be presented in support of such a view other than a theoretical consideration of the thinning necessary for failure of the formation.

Solution of this sort would result in a concentration of the residual minerals of the Oneota on the top layer of the underlying sandstone. This fact was not considered at the time the various series of samples were collected, so that this peculiarity of mineral distribution is shown in only one series, No. 4, from Ottawa. At several localities—Stillwater, Ottawa, St. Peter, and Mankato—and at numerous places noted by Winchell and Upham, concentrations of iron oxide and clay, ranging in thickness from a fraction of an inch to several inches, occur at the contact and along the walls of the numerous joints. This concentration of iron oxide suggests circulation of ground water as well as solution of the dolomite along these joints, thus forming the cavities already discussed. It is not thought probable that subaërial erosion produced the clay-iron oxide zone for reasons previously discussed.

Below are three sections, studied in detail both in the field and microscopically. Other sections were measured and studied, but are not here reported in complete form.

RED WING, MINNESOTA. (BARN BLUFF) SERIES 3

Formation	Thickness	Description
Oneota	To Top of Bluff	Massive buff dolomite
	2 ft. 4 in.	Thin-bedded, fine-grained, friable white sandstone
	5 ft. 10 in.	Massive buff dolomitic sandstone
	8 ft. 4 in.	Thin-bedded, fine-grained, dolomitic sandstone with brown clay pebbles
	2 ft. 6 in.	Friable, buff, fine-grained sandstone with brown clay pebbles
	4 ft. 10 in.	Fine-grained dolomitic sandstone
	3 ft. 5 in.	White, friable, massive sandstone
	2 ft. 10 in.	Sandstone breccia
Jordan sandstone	7 in.	Fine-grained, white, friable sandstone
	7 in.	Fine-grained, white, friable sandstone with green shale partings
	3 ft. 5 in.	Medium-grained, friable, white sandstone
	3 ft. 7 in.	Thin-bedded, buff, dolomitic sandstone
	11 ft. 8 in.	Massive, coarse, cross-bedded white to buff sandstone, dolomitic toward top
	6 ft. 2 in.	Massive, coarse, friable, buff sandstone with short clay lenses in upper part

SUGAR LOAF, WINONA, MINNESOTA. SERIES 8

Formation	Thickness	Description
Oneota	50 ft.	Massive, buff dolomite with cellular zones to top of knob
	7 ft. 6 in.	Massive, buff dolomite
	1 ft. 0 in.	Buff dolomite with shells
	0 ft. 10 in.	Dolomite with green and pink streaks
	0 ft. 9 in.	Cryptozoan zone
	5 ft. 0 in.	Dolomite, somewhat oolitic with secondary calcite veins and crystals
	1 ft. 6 in.	Oolitic, sandy dolomite with small sandstone pebbles
	1 ft. 10 in.	Sandstone conglomerate
	1 ft. 3 in.	Brecciated zone in massive white sandstone
Jordan sandstone	1 ft. 11 in.	Cross-bedded, white, nodular sandstone

LANESBORO, MINNESOTA. (ROAD CUT ON HIGHWAY 1 MILE EAST OF TOWN.) SERIES 9

Formation	Thickness	Description
Oneota	1 ft. 1 in.	Cellular dolomite
	1 ft. 9 in.	Fine-grained, gray, to light gray, even-bedded dolomite
	1 ft. 8 in.	Fine-grained, buff dolomite
	2 ft. 8 in.	Cellular, gray dolomite
	2 ft. 2 in.	Drab dolomite
	2 ft. 0 in.	Massive, sandy dolomite or dolomitic sandstone
Jordan sandstone	2 ft. 9 in.	Friable, buff, cross-bedded sandstone
	5 ft. 3 in.	Hard, massive, sandstone with clay pebbles to ditch

METHOD OF INVESTIGATION

Numerous sections showing the contact between the Jordan and Oneota were studied and sampled in detail. The samples were taken at short vertical intervals to show the changes in lithology from a few feet below the contact to a few feet above it. The actual distance between samples ranged from a few inches to several feet, depending on the uniformity of the particular bed sampled.

The method used in preparing and mounting the heavy mineral separates for study has been described by Graham.[8] The minerals found in the heavy concentrates are garnet, tourmaline, zircon, staurolite, rutile, brookite, anatase, cassiterite, corundum, augite, kyanite, epidote, topaz, leucoxene, and several opaque minerals such as magnetite, pyrite, ilmenite, *et cetera*. The opaque minerals were not used in the present study for correlative purposes, because of the difficulty of determining them definitely in the permanent mounts.

The minerals found in the slides were studied with reference to total percentage, shape, color, inclusions, surface features, percentage of grain sizes in millimeters, alteration, and any other special features that might be discovered in the course of the investigation. The percentages were determined on a count of two hundred and fifty grains in all, but three or four slides were made from samples containing less than this number of heavy mineral grains.

[8] William A. P. Graham, "A Textural and Petrographic Study of the Cambrian Sandstones of Minnesota," *Jour. Geol.*, Vol. XXXVIII (1930), p. 704.

PETROGRAPHIC RESULTS

Series 1, Stillwater, Minnesota.—The contact at Stillwater is almost knife edge in sharpness and occurs between samples 2 and 3. The analysis of the microscopic data in Table I usually shows the break between samples 2 and 3 also. However, in analyzing the various changes in the microscopically studied features for each mineral, it is found that marked breaks occur either above or below the contact as determined in the field.

Of the thirty-one features studied in this series, nine indicate the contact occurs between samples 2 and 3, one places it between 1 and and 2, and two between 3 and 5. The remaining nineteen points are of no value in locating the contact, since they do not show breaks greater than those found in a single formation.

Evidence of erosion of new rock types is found in the change in percentage of inclusions and color observed in the garnet grains; change in color and inclusions in the tourmaline grains; and the change in the percentage of inclusions in the zircon grains. These changes are not great, thus suggesting the exposure of relatively small areas of rocks different from those exposed during the Cambrian period.

Series 4, Ottawa, Minnesota.—The contact between the Jordan and Oneota at Ottawa is very sharp, although somewhat undulatory. On the basis of field evidence the contact occurs between samples 2 and 3. The microscopic study usually suggests a break at the same place, although the bulk of the evidence is rather unsatisfactory.

Forty-one features were studied in these samples; five indicate the contact is between samples 1 and 2, seven locate it between samples 2 and 3, and twenty-nine are of no significance in locating the contact.

Sample 2 was taken immediately below the dolomite, where solution of the dolomite would result in a concentration of the relatively insoluble mineral matter derived from its solution. A high concentration of iron oxide at the contact also suggests solution of the dolomite. Contamination of the sample in this way may have occurred giving a mixture of residual minerals of the dolomite on the upper layer of the Jordan. The actual contact is a short distance below the dolomite, at the base of a silt clay zone composed of the accumulated residual material.

TABLE I

This table shows the complete analysis of the heavy mineral separates from the several samples. The percentage column shows the percentage of each mineral in the specimen, the other features individually total 100 per cent for every mineral reported. The following specimens are from the Jordan sandstone: 1–1 and 2, 4–1 and 2, 6–1, 7–1, 8–1, 9–1, 2, and 3; all other samples are from the Oneota formation.

	GARNET														
		COLOR			INCL.		SHAPE				SIZE				
SERIES NO. SAMPLE NO.	PERCENTAGE	Pink	Blue	Colorless	Without	With	Round	Sub-round	Sub-angular	Angular	½ mm.	¼ mm.	⅛ mm.	1/16 mm.	1/16–1/8 mm.
9–13	84	1	99	.....	93	7	.....	.....	.....	100	.....	.....	.....	13	87
9–11	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
9–9	76	.....	100	.....	87	13	.....	.....	.....	100	.....	.....	.....	17	83
9–7	73	.....	100	.....	97	3	.....	.....	2	98	.....	.....	.....	3	97
9–6	88	15	85	.....	65	35	3	77	14	6	.....	.....	46	54	.....
9–5	57	15	85	.....	50	50	4	65	29	2	.....	1	25	73	1
9–4	80	.....	100	.....	75	25	3	33	55	9	.....	.....	6	84	10
9–3	55	.....	100	.....	66	34	.....	38	44	20	.....	.....	25	66	9
9–2	69	4	96	.....	85	15	1	28	47	24	.....	.....	3	94	3
9–1	77	12	88	.....	76	24	2	32	37	29	.....	.....	6	84	10
8–7	32	.....	78	22	71	29	.....	17	35	48	.....	.....	.....	65	35
8–6	39	.....	86	14	54	46	.....	15	54	31	.....	.....	.....	94	6
8–5	66	.....	96	4	69	31	.....	15	67	18	.....	.....	4	87	9
8–4	61	5	95	.....	78	22	3	14	56	27	.....	.....	40	54	6
8–3	69	1	98	1	74	26	.....	13	45	42	.....	1	14	60	25
8–2	83	58	42	.....	59	41	7	34	47	12	2	23	54	21	.....
8–1	82	20	80	.....	63	37	.....	26	29	45	.....	1	49	44	6
7–6	36	.....	100	.....	50	50	.....	18	27	55	.....	.....	.....	80	20
7–4	66	.....	100	.....	83	17	.....	.....	17	89	.....	.....	.....	8	92
7–2	62	.....	100	.....	91	9	.....	.....	.....	100	.....	.....	.....	19	81
7–1	68	3	97	.....	75	25	.....	2	38	60	.....	.....	5	46	59
6–3	50	.....	100	.....	86	14	.....	5	27	68	.....	.....	.....	4	96
6–2	66	.....	100	.....	69	31	.....	.....	14	86	.....	.....	1	11	88
6–1	8	.....	100	.....	88	12	.....	.....	38	72	.....	.....	.....	12	88
4–5	18	.....	100	.....	50	50	.....	50	50	.....	.....	.....	.....	50	50
4–4	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
4–3	4	.....	100	.....	50	50	.....	.....	100	.....	.....	.....	.....	100	.....
4–2	2	100	.....	.....	50	50	50	50	.....	.....	.....	.....	.....	100	.....
4–1	39	66	34	.....	34	66	25	33	8	33	.....	8	34	16	42
1–6	15	.....	100	.....	100	.....	.....	.....	.....	100	.....	.....	.....	.....	100
1–5	27	.....	100	.....	55	45	.....	.....	33	67	.....	.....	.....	22	78
1–3	77	.....	100	.....	91	9	.....	.....	7	93	.....	.....	.....	3	97
1–2	72	36	64	.....	62	38	4	16	33	47	.....	2	29	69	.....
1–1	40	35	58	7	51	49	6	16	3	75	.....	5	23	67	5

TABLE I—*Continued*

Series No. Sample No.	Percentage	Tourmaline																
		Color					Incl.		Shape				Size					
		Brown	Green	Blue	Black	Colorless	Without	With	Round	Sub-round	Sub-angular	Angular	½ mm.	¼ mm.	⅛ mm.	1/16 mm.	0–1/16 mm.	
9–13	3	67	33	....	....	....	....	100	....	....	33	67	....	....	....	67	33	
9–11	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	
9–9	6	33	50	17	....	....	....	100	....	....	17	83	....	....	....	....	100	
9–7	11	50	50	....	....	....	50	50	....	....	25	75	....	....	....	....	100	
9–6	4	70	15	15	....	....	15	85	37	63	....	....	....	....	24	76	....	
9–5	28	53	37	10	....	....	27	73	26	74	....	....	....	....	44	56	....	
9–4	7	100	....	....	....	....	15	85	25	63	....	12	....	....	12	51	37	
9–3	40	51	46	3	....	....	31	69	15	70	15	....	....	....	58	42	....	
9–2	19	40	60	....	....	....	50	50	30	65	5	....	....	....	12	88	....	
9–1	9	74	26	....	....	....	20	80	10	90	....	....	....	....	20	74	6	
8–7	29	87	9	4	....	....	26	74	....	17	4	79	....	....	....	13	87	
8–6	51	86	13	....	....	....	50	50	....	18	7	75	....	....	2	16	82	
8–5	20	86	14	....	....	....	12	88	....	50	....	50	....	....	7	43	50	
8–4	32	78	17	3	2	....	10	90	13	48	10	29	....	....	15	56	29	
8–3	19	88	8	4	....	....	30	70	8	20	....	72	....	....	16	12	72	
8–2	13	77	16	....	7	....	84	16	15	15	....	70	....	....	15	15	70	
8–1	7	56	33	....	....	11	....	100	23	44	....	33	....	....	33	21	46	
7–6	19	25	75	....	....	....	....	100	....	25	....	75	....	....	....	25	75	
7–4	7	62	30	8	....	....	....	100	....	8	3	61	....	....	....	25	75	
7–2	23	50	50	....	....	....	....	100	4	4	13	79	....	....	....	30	70	
7–1	13	46	46	8	....	....	....	100	23	46	31	....	....	....	....	38	38	24
6–3	14	100	....	....	....	....	40	60	....	....	....	100	....	....	....	....	100	
6–2	14	58	42	....	....	....	14	86	....	42	58	....	....	....	....	14	43	43
6–1	7	12	76	....	....	12	18	82	....	44	23	23	....	....	....	....	12	88
4–5	18	100	....	....	....	....	....	100	....	50	....	50	....	....	....	50	50	
4–4	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	
4–3	4	100	....	....	....	....	....	100	....	50	....	50	....	....	....	....	100	
4–2	26	76	21	....	....	3	15	85	24	45	21	10	....	....	17	73	10	
4–1	51	87	13	....	....	....	....	100	67	33	....	....	....	....	36	58	6	
1–6	44	93	....	7	....	....	50	50	....	....	17	83	....	....	....	....	100	
1–5	33	100	....	....	....	....	55	45	....	....	17	83	....	....	....	9	91	
1–3	3	67	33	....	....	....	33	67	....	33	33	34	....	....	....	....	100	
1–2	16	62	35	3	....	....	6	94	12	64	24	....	....	5	30	60	5	
1–1	38	77	14	....	8	....	16	84	14	21	9	56	....	....	13	51	36	

TABLE I—*Continued*

Series No. Sample No.	Percentage	Zircon																
		Color			Zoned		Incl.		Shape				Size					
		Pink	Blue	Colorless	Without	With	Without	With	Round	Sub-round	Sub-angular	Angular	¼ mm.	¼ mm.	¼ mm.	⅛ mm.	⅛ mm.
9 13	8	70	30	....	70	30	88	12	....	....	....	100	....	....	....	12	88
9–11	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....	....
9–9	7	100	....	....	....	....	100	....	....	....	....	100	....	....	....	....	100
9–7	8	83	17	....	83	17	100	....	....	....	33	67	....	....	....	....	100
9–6	8	94	6	....	94	6	70	30	17	61	22	....	....	....	6	82	12
9–5	15	100	....	....	....	....	28	72	14	79	7	....	....	....	....	75	25
9–4	10	78	22	....	78	22	75	25	19	75	....	6	....	....	....	78	22
9–3	5	100	....	....	100	....	100	....	....	....	75	25	....	....	....	25	75
9–2	11	71	20	9	80	20	85	15	....	55	45	....	....	....	....	67	33
9–1	10	100	....	....	100	....	100	....	....	60	40	....	....	....	....	80	20
8–7	36	62	34	4	66	34	85	15	....	33	56	11	....	....	3	40	57
8–6	10	56	44	....	56	44	67	33	....	56	33	11	....	....	11	67	22
8–5	10	50	50	....	50	50	70	30	....	63	25	12	....	....	....	63	37
8–4	7	64	36	....	64	36	50	50	7	57	36	....	....	....	....	71	29
8–3	9	80	20	....	80	20	53	47	7	60	20	13	....	....	....	40	60
8–2	4	100	....	....	100	....	75	25	25	75	....	....	25	....	25	50	....
8–1	3	50	25	25	75	25	75	25	....	50	50	....	....	....	25	25	50
7–6	26	83	17	....	83	17	83	17	....	17	50	33	....	....	....	....	100
7–4	26	100	....	....	100	....	80	20	....	10	61	29	....	....	....	4	96
7–2	12	86	14	....	86	14	100	....	....	....	14	86	....	....	....	7	93
7–1	16	70	30	....	70	30	67	33	....	17	61	22	....	....	....	33	67
6–3	11	75	25	....	75	25	100	....	....	25	25	50	....	....	....	....	100
6–2	20	87	13	....	87	13	67	33	....	7	40	53	....	....	....	....	100
6–1	84	100	....	....	100	....	91	9	....	7	43	50	....	....	....	....	100
4–5	38	100	....	....	100	....	100	....	....	50	50	....	....	....	....	....	100
4–4	15	100	....	....	100	....	....	100	....	100	....	....	....	....	....	100	....
4–3	90	80	20	....	82	18	72	28	....	1	28	71	....	....	....	....	100
4–2	51	98	2	....	92	8	57	43	3	53	38	6	....	....	3	45	52
4–1	10	100	....	....	100	....	67	33	67	33	....	....	....	....	....	100	....
1–6	15	100	....	....	100	....	75	25	....	....	75	25	....	....	....	25	75
1–5	8	50	50	....	50	50	87	13	....	....	75	25	....	....	....	13	87
1–3	15	94	6	....	94	6	94	6	....	....	50	50	....	....	....	....	100
1–2	12	93	7	....	93	7	39	61	....	75	25	....	....	....	....	85	15
1–1	22	7	93	....	77	23	51	49	9	....	13	78	....	....	15	57	28

TABLE I—*Continued*

Series No. Sample No.	Staurolite												
	Percentage	Color	Incl.		Shape				Size				
		Blue	Without	With	Round	Sub-round	Sub-angular	Angular	½ mm.	¼ mm.	⅛ mm.	1/16 mm.	0–1/16 mm.
9–13	1	100	.....	100	.....	.....	.....	100	.....	.....	.....	100	.....
9–11	100	100	.....	100	.....	.....	.....	100	.....	.....	.....	82	18
9–9	8	100	.....	100	.....	.....	.....	100	.....	.....	.....	33	67
9–7	4	100	25	75	.....	.....	25	75	.....	.....	.....	75	25
9–6	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
9–5	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
9–4	1	100	.....	100	.....	.....	.....	100	.....	.....	.....	.....	100
9–3	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
9–2	1	100	.....	100	.....	.....	.....	100	.....	.....	.....	100	.....
9–1	3	100	.....	100	.....	.....	100	.....	.....	.....	.....	100	.....
8–7	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
8–6	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
8–5	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
8–4	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
8–3	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
8–2	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
8–1	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
7–6	19	100	25	75	.....	.....	.....	100	.....	.....	.....	25	75
7–4	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
7–2	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
7–1	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
6–3	11	100	.....	100	.....	.....	25	75	.....	.....	.....	100	.....
6–2	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
6–1	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
4–5	26	100	.....	100	.....	.....	33	67	.....	.....	.....	67	33
4–4	85	100	.....	100	.....	.....	.....	100	.....	.....	.....	.....	100
4–3	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
4–2	18	100	19	81	.....	25	45	30	.....	5	22	73	.....
4–1	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
1–6	22	100	83	17	.....	.....	.....	100	.....	.....	.....	.....	100
1–5	32	100	18	82	.....	.....	82	18	.....	.....	.....	36	64
1–3	5	100	?	?	.....	.....	20	80	.....	.....	.....	.....	100
1–2	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
1–1	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....

TABLE I—*Continued*

Series No. Sample No.	Other Minerals* Percentage	Incl. Without	Incl. With	Shape Round	Shape Sub-round	Shape Sub-angular	Shape Angular	Size ½ mm.	Size ¼ mm.	Size ⅛ mm.	Size 1/16 mm.	Size 1/32 mm.
9–13	4	100					100					100
9–11												
9–9	3	100				75	25				75	25
9–7	4	75	25			50	50			25		75
9–6												
9–5												
9–4	2	100				100					100	
9–3												
9–2												
9–1	1	75	25				100				100	
8–7	3	100				67	33					100
8–6												
8–5	4	100					100					100
8–4												
8–3	3	80	20			20	80			20		80
8–2												
8–1	8	40	60			20	80					100
7–6												
7–4	1	100					100					100
7–2	3	100					100					100
7–1	3	100					100					100
6–3	14	80	20		35	15	50				20	80
6–2												
6–1	1	100					100					100
4–5												
4–4												
4–3	2	100				100						100
4–2	3	100				67	33				33	67
4–1												
1–6	4	100					100					100
1–5												
1–3												
1–2												
1–1												

* Minerals summarized in this column are Rutile, Anatase, Brookite, Cassiterite, Corundum, Augite, Epidote, Kyanite, and Topaz.

The evidence for the erosion of new rock types during early Ordovician time is not clearly shown in this section. However, the change in the percentage of garnet and tourmaline grains having inclusions and the change in the color of the tourmaline are suggestive of erosion of new rock units on a very moderate scale.

Series 6, Mankato, Minnesota.—The contact between the Jordan and Oneota in the field is very sharp, but irregular, and occurs between samples 1 and 2. (See Fig. 2.) The evidence presented by the microscopic examination of the specimens is quite indefinite and cannot be used without care in locating the contact.

From a total of fifty-one points studied microscopically, five indicate the contact lies between samples 1 and 2, three indicate it to be between samples 2 and 3, and forty-three are either indefinite or of little value because of the extreme rarity of certain minerals or their occurrence in both formations in one or more sections under almost identical conditions.

The evidence supporting erosion of new rock types during Oneota time is not conclusive but is suggestive. The break in color of both tourmaline and zircon and the change in percentage of the inclusions in zircon suggest erosion of new rock types on a small scale, but not necessarily a break in sedimentation. The field evidence, however, indicates a break in sedimentation. No brecciation or re-working of the Jordan is exposed opposite Sibley Park, Mankato.

Series 7, Minneiska, Minnesota.—The contact in this section cannot be accurately located either in the field or microscopically. The upper portion of the Jordan has been partially re-worked by the Oneota sea, as evidenced by the gradation of the sandstone into the dolomite. Sand from the underlying sandstone, as well as from bordering areas, was in a state of agitation during the deposition of the basal Oneota beds, resulting in their unusually sandy nature. Considerable secondary carbonate has been deposited for some distance below the contact, tending to mask the contact even more.

The probable location of the contact on the basis of field evidence is between samples 1 and 2, while on the basis of microscopic evidence it is between 4 and 6. In this case field evidence is given preference in locating the contact, since at any contact where re-working of the underlying formation has taken place during a later sea ad-

vance, a gradation in the mineral concentrations of the two formations will result, making the break gradual, the thickness of the gradation zone being proportional to the degree of erosion.

Series 8, Sugar Loaf, Winona, Minnesota.—The contact at Sugar Loaf, on field evidence, is placed between samples 1 and 2, while on the basis of microscopic evidence it falls between 5 and 6. The section shows a brecciated zone immediately above the cross-bedded, nodular, white sandstone (see section). The breccia fragments are of this same white sandstone. Above the breccia zone is a conglomerate which grades upward into a sandy dolomite containing small sandstone pebbles. These three zones of re-worked Jordan sandstone are 4 feet 7 inches thick and occur directly below a somewhat oolitic dolomite bed. The evidence for the re-working of the Upper Jordan beds by the advancing Oneota sea is conclusive and does not permit the placing of the contact at a higher horizon in spite of the microscopic evidence.

Of fifty-six features studied in this section with reference to the different minerals in each slide, four indicate that the contact is between samples 5 and 6, one locates it between 6 and 7, and fifty-one are of no value in differentiating these two formations.

Erosion of new rock types is not clearly indicated by the minerals found. The change in percentage of the various colors of garnet is the only feature observed which is indicative of exposure of new formations to erosion.

Series 9, Lanesboro, Minnesota.—The contact, on the basis of field evidence, is located between samples 3 and 4 at the contact of the 2 foot 9 inch friable, buff, cross-bedded sandstone and the 2 foot massive sandy dolomite bed (see section). The microscopic evidence is not at all consistent; various features locate the contact either between samples 3 and 4, or 6 and 7, at the base, middle, or top of the massive sandy dolomite layers. The microscopic evidence shows this bed to be a mixture of re-worked Jordan sandstone and carbonate.

Out of forty-six features studied in this series, one locates the contact at the proper point, between samples 3 and 4, two between 5 and 6, and four between 6 and 7, the other thirty-nine features give no hint as to the location of the contact.

No evidence for erosion of new rock types was found in the rocks from this series. Several variations occur in the minerals at the contact, but similar changes both above and below nullify their value for interpretation.

In several of the sections studied both in the field and microscopically the evidence used in locating the contact by the two methods gives divergent results. In cases of disagreement the contact determined by microscopic evidence is always higher than when determined on field evidence, except in those cases where the breccia and silty clays are absent and solution of the dolomite has occurred. This is what one would expect to find at an unconformable contact if erosion of the underlying formation took place during the advance of a later sea. The lowest disturbed beds would have dominantly the mineral composition of the eroded formation. A gradual transition in this mineral composition would take place as the top of the disturbed or re-worked zone is approached, resulting from mingling of the minerals characteristic of the overlying formation with those of the lower in increasing proportion. When this transition zone is passed, the minerals characteristic of the top bed only will prevail. It is suggested that the zone in which the Jordan and Oneota minerals occur together may actually extend above the zone of disturbance, owing to inwash of typical Jordan minerals from bordering land areas where the Jordan sandstone was still exposed to erosion. The great mineralogical similarity of these two formations is best explained in this way.

E. O. Ulrich,[9] in discussing his new classification of Paleozoic formations in Wisconsin, notes a large unconformity between the lowest Ozarkian and St. Croixan formations in both the eastern and western portions of the state. He concludes that the Middle and Lower Ozarkian formations are essentially absent in this area, whereas in Missouri they are represented by approximately 560 feet of dolomite. During this interval erosion of the Jordan sandstone must have taken place, with the removal of a considerable thickness of rock. The Jordan of the present time is a very friable formation and easily eroded by present-day streams. If such a long period of time elapsed, as is postulated by Ulrich, to allow for the accumulation of

[9] "Notes on New Names in Table of Formations, etc.," *op. cit.*, pp. 82–83, 105.

approximately 560 feet of dolomite in Missouri, one would expect the Jordan to show a much greater variation in thickness below the Oneota than is found; and new rock types should have been exposed to erosion, resulting in new minerals in the re-worked portion of the Jordan as well as in the dolomite. The evidence already stated indicates but a relatively short erosion interval. It might be suggested that the land area during the erosion interval following Jordan time was essentially without relief and very near sea level. It is questionable if a land area even of such low relief would not develop recognizable stream channels during an erosion interval as long as that postulated by Ulrich. This would be especially true of that portion of the area farthest removed from the point of deposition. The area under consideration must have been at least several hundred miles from the early Ozarkian sea.

Dr. C. R. Stauffer[10] has found the Jordan to thicken from 68 to 70 feet at Stillwater, on the northern edge of the area considered, to 117–28 feet at Hokah, in southeastern Minnesota—an air-line distance of about 115 miles. The thickness of the Jordan between these two locations varies between the two thicknesses stated, largely due to the difficulty in locating the top and bottom of the formation. In each section considered the Jordan is capped by the Oneota.

Considerable evidence in substantiation of the presence of a large diastem[11] at this contact has been found in the field: (1) the presence of a brecciated zone at the top of the Jordan sandstone; (2) the presence of post-Jordan fossils just above the brecciated zone; (3) absence of notable relief at the top of the Jordan; (4) the petrographic study indicates but a relatively short period of erosion between the deposition of these two formations; (*a*) the mineral associations found in the Jordan and Oneota are remarkably similar, (*b*) a long erosion interval would have resulted in the removal of the very fri-

[10] Personal communication.

[11] Some question exists in the writer's mind as to the desirability of using the term "large diastem," for "small unconformity" to indicate the magnitude of this break in sedimentation. The term "diastem" is normally used to denote a break in sedimentation during which time a deposit of formational proportions was not laid down elsewhere. A "small unconformity" would allow the deposition of a formation at some other place during the erosion interval. Additional evidence is considered necessary before it is possible to determine with certainty which of these two terms should be used.

able Jordan sandstone from a considerable territory, exposing relatively large areas of different rock types, making the mineralogical differences between the two formations easily recognizable.

SUMMARY

Of the various heavy minerals found in the samples examined, garnet, tourmaline, and zircon are the most persistent species. They occur in both formations in varying amounts, and sometimes can be used in determining the contact. Of the various features of each mineral studied, the percentage of the mineral in each sample, the change in the percentage of grain colors, and the variation in the precentage of inclusions are of most value in determining these two formations. Shape and size of the individual grains are not of great utility, since there is a definite relationship between the degree of rounding and size of the individual grains.

Solution of the Oneota at the contact has resulted in a concentration of the residual Oneota minerals at the top of the Jordan sandstone. The thickness of such a residual accumulation is proportional to the amount of dolomite dissolved and the concentration of relatively insoluble original material in the dolomite. The solution of the Oneota at the contact with the Jordan is supported by field evidence.

A gradation of Jordan minerals into the basal portion of the Oneota is common.

A break in sedimentation between the Oneota and Jordan is indicated by both field and miscroscopic evidence.

Petrographic means of locating the contact between these two formations are not consistently of value, owing to the re-working of the Jordan sandstone by the advancing Oneota sea and to the slight mineralogic differences found in the two formations.

THE INTRUSIVE RELATIONS OF THE REPUBLIC GRANITE

CARL A. LAMEY
Northwestern University

ABSTRACT

The larger part of the pre-Cambrian rocks of northern Michigan known as the Southern Complex, and previously designated Archean, is thought to be composed of a post–Upper Huronian granite which has intruded and extensively metamorphosed the Huronian sediments and which probably extends westward some distance beneath the Upper Huronian rocks in the vicinity of Lake Michigamme. Because of its excellent development in the vicinity of Republic, it is suggested that this granite be designated the "Republic granite." Structural relations of this granite indicate that it originated under a regional tectonic stress directed southwestward.

INTRODUCTION

The territory under consideration comprises a somewhat rectangular area in the northern peninsula of Michigan. Its northern boundary extends generally westward from Marquette to the west side of Lake Michigamme, and its southern limit is near the Michigan-Wisconsin boundary, as shown by the accompanying index map (Fig. 1).

FIG. 1.—Index map showing location of area described.

A variety of topographic forms exists within the area. In general it may be said that the marginal parts are rough and hilly. This roughness around the margins is due to ridges composed of metamorphosed sediments, to somewhat rounded granite hills which rise knoblike above the surrounding country, to larger hills composed of diorite, and to glacial outwash and moraines with well-developed knob and kettle topog-

raphy. Toward the interior of the area the bedrock is almost everywhere obscured by glacial outwash, and the diversity of topographic forms characteristic of the margins is lacking. This central sandy area of little relief is known locally as the Sand Plains.

Geologically the area is a part of the general Lake Superior pre-Cambrian region and contains a variety of rock types. The igneous rocks consist of granite, diorite, diabase, and greenstones, while the

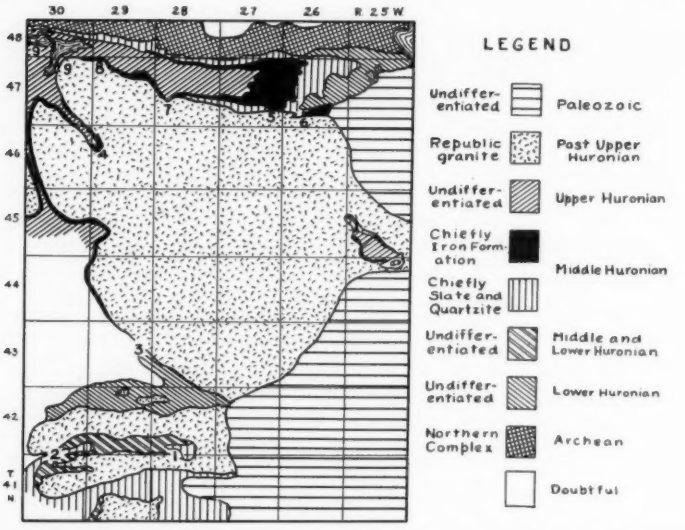

FIG. 2.—Map showing the general geology of the area described. After Allen and Barrett, modified.

sedimentary and metamorphic rocks are principally the complexly folded Huronian sediments consisting of conglomerates, quartzites, graywackes, slates, dolomites, and iron formation, although some Paleozoic sediments are present. Practically the entire central part of the area under consideration is occupied by granite. The Huronian sediments occur around the margins of this granite and are especially well developed in the Marquette district, while the Paleozoics are confined to the eastern and southeastern parts of the area (Fig. 2). The diorite forms knobs and sometimes ridges in the marginal parts of the area.

In many places the Huronian sediments are cut by granite dikes,[1] are intruded by the granite of the central part of the area, and show unmistakable evidences of contact metamorphism. It was these relations that led the writer to suggest that the granite was probably the source of the dikes and the cause of the contact metamorphism, and that it is likely to be post–Upper Huronian rather than Archean.[2] In addition to the work done previously the entire summer of 1931 was devoted to a more thorough and more extensive field study of this problem. The results obtained indicate without doubt that the granite intrudes Lower, Middle, and Upper Huronian formations. The purpose of this paper is to set forth the proofs of this relation and to suggest a name for the granite, which has hitherto been known as a part of the Southern Complex. A description of the metamorphism produced by the Republic granite has not yet been completed. About one hundred thin sections of metamorphosed sediments from various parts of the granite periphery have been examined and a separate paper dealing with the metamorphic phase of the intrusive relations is being prepared.

THE GRANITE

In general, the granite is pink to red in color and is coarsely granular and porphyritic, but it has a border facies which is often gray to white and finely granular, and a pegmatitic facies which is much coarser than the main granite mass. The feldspar phenocrysts, many of which show Carlsbad twinning, are frequently 2 inches long and $\frac{1}{2}$ inch across. Quartz is present in relatively small crystals as a rule, and both muscovite and biotite are present, although mica is not an abundant constituent. Thin sections show that the predominant feldspar is microcline, but that micro-perthite and an acid plagioclase are also present. The abundance and identity of all the accessory minerals are not yet fully known. However, from partially completed studies of the accessory minerals recovered as heavy constituents from crushed specimens it is known that molybdenite, pyrite, pyrrhotite, chalcopyrite, magnetite, tourmaline, garnet, apatite, fluorite, zircon, and ilmenite are present. Some of these—molyb-

[1] C. A. Lamey, "Granite Intrusions in the Huronian Formations of Northern Michigan," *Jour. Geol.*, Vol. XXXIX (1931), pp. 288–95.

[2] *Ibid.*, p. 295.

denite, for example—occur in crystals ½ inch across in a pegmatitic facies of the granite.

The pegmatitic facies of the granite has two modes of occurrence: (1) as dikes cutting the main granite without sharp contacts and (2) as gradational areas within the main mass. Some of these latter areas are roughly circular with a diameter of 5–50 feet or more.

Around the periphery of the main mass the granite becomes much finer grained, frequently changes to a grayish color, especially in the vicinity of quartzites, and often assumes a gneissoid character. In places the border phase of the granite is separated from the Huronian sediments by an area containing no outcrops, but in many places it is possible to trace the connection between the granite and the sediments, and it is then found that the granite passes from the normal type to the more gneissoid type, then to material that would be classed as gneiss, and finally into the Huronian sediments, showing clearly that this gneissoid border phase is a result of contact metamorphism.

CONTACT RELATIONS

The contact relations existing between the granite and the Huronian sediments change to some extent from place to place, but the essential features of these relations are (1) pronounced contact metamorphism; (2) *lit-par-lit* injections and gneissoid gradations; (3) discordant contacts, the granite cutting across the strike of the sediments; and (4) the occurrence of granite dikes in the sediments. The relations in a few typical areas will be described briefly, although many other instances could be cited. The locations of these areas are shown on the geological map (Fig. 2), each area being designated by a number. Attention has been called to a number of the granite dikes cutting the sediments,[3] and those descriptions will not be repeated here.

1. *The Felch area.*—About ½ mile southeast of Felch is a marble quarry in the Lower Huronian Randville dolomite. The dolomite is well exposed in the western part of the quarry where it forms a closely folded syncline striking slightly north of west and plunging southeast. The material in the eastern part of the quarry is an amphibole schist with schistosity striking approximately the same as the axis of

[3] *Ibid.*

the syncline. It extends westward toward the dolomite, and in some places it apparently overlies and is infolded with the dolomite. Much of the Randville dolomite is greatly metamorphosed and in places contains large crystals of tremolite and an amphibole similar to actinolite, pyroxene, phlogopite, and smaller amounts of chalcopyrite, pyrrhotite, and pyrite. Because of the intensity and character of the metamorphism, it was thought that the granite contact must be very close to the quarry. This proved to be the case, and the granite surrounds the entire southeastern part and cuts off the amphibole schist (Fig. 3). Small dikes from the main granite project northward into the schist and toward the dolomite, and either these or a part of the main granite apparently intrude the dolomite from beneath, as indicated by large quantities of quartz and a concentration of sulphides in some of the more intensely metamorphosed spots.

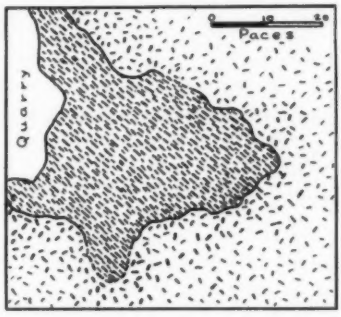

FIG. 3.—Map showing the Republic granite cutting discordantly across amphibole schist (probably Lower Huronian) near Felch.

2. *The Randville area.*—A similar but much larger quarry is found a short distance east of Randville, close to the junction of state roads Nos. 45 and 69, and easily accessible from the latter highway. The Randville dolomite again shows evidence of extensive contact metamorphism, but the amphibole schist is not present in this quarry. Granite was known to occur about $\frac{1}{2}$ mile north of the quarry, but no granite was found closer to it until late in the summer of 1931 when road excavations on state road No. 45 uncovered granite close to the intersection of the two highways previously mentioned, and within a few hundred feet of the quarry. Although the actual contact between the Randville dolomite and the granite cannot be observed, there seems little reason to doubt that the former is intruded by the latter.

3. *The Turner area.*—This area, approximately southwest of Turner $\frac{3}{4}$ of a mile, is not easily accessible and the geology is not well

known. Outcrops which stand up as ridges consist of sediments very thoroughly intruded with granite, and in places large patches of sediment appear to have been caught up and surrounded with granite. The sediments apparently are part of the Lower Huronian Sturgeon quartzite and Randville dolomite series, as both quartzite and dolomite can be recognized in places and also rocks of these series are known to occur farther southeastward in the Sturgeon River trough. In general in this vicinity, to the southeast the sediments of the Sturgeon quartzite and Randville dolomite series are more abundant and can be traced some distance from the granite, whereas to the northwestward the granite is found close to the Vulcan iron formation. Apparently the granite is cutting discordantly into the northern limb of the Sturgeon River syncline, across the Sturgeon quartzite and Randville dolomite.

4. *The Republic area.*—As the Middle Huronian iron formation and the underlying Ajibik quartzite are traced around the south end of the Republic trough, following the southernmost part of the outcrops, a striking amount of contact metamorphism is observed. In places highly garnetiferous and micaceous bands alternate with grüneritic bands, whereas in other places, where the Ajibik quartzite contained argillaceous material, bands containing much andalusite are found. Southward from these outcrops, in a marshy area, granite is found at distances ranging from 75 to 200 feet. The best example of the actual contact between the granite and the Ajibik quartzite occurs in the northeastern part of Section 18, where the quartzite and the iron formation strike approximately N. 20° E. and the granite crops out about 75 feet south of the most southern quartzite exposure. Here, then, as nearly as can be ascertained in the field at this location, the quartzite strikes directly into the granite. Of more significance, however, is the fact that on the western side of this granite outcrop, which forms a hill about 50 feet high, the granite becomes gneissoid and patches of undoubted sedimentary origin are found incorporated in it, establishing the actual intrusive contact between granite and sediments.

If the Huronian formations are followed westward and northwestward from here along the strike, which changes to northwest, the dump pile of the abandoned West Republic mine will be found on the east bank of the Michigamme River. Examination of the

material on the dump pile shows that the iron-bearing part consists chiefly of specularite and magnetite, and that not infrequently the iron formation contains pyroxene crystals and dikelike veins of quartz and feldspar. Across the river, in the line of strike, the iron formation and quartzite are missing, and in their place there appears a large hill of granite. Though there is no positive evidence that the granite has cut discordantly across the strike of the Negaunee iron-bearing formation, the field relations suggest very strongly that such is the case.

5. *The Lake Palmer area.*—In this area there are many places that show clearly that the granite intrudes the Middle Huronian formations, which have a general strike of about N. 80° W. and dip northward. Southward across the strike of the Siamo slate and Ajibik quartzite in the northwestern part of Section 26, about 1 mile west of Lake Palmer, several granite dikes are found cutting the Siamo slate. Farther southward the Ajibik quartzite is encountered and into this several domelike masses of granite have been intruded. One of these masses is about 130 feet across. In general, the masses increase in size until finally the quartzite disappears and nothing but granite and gneissoid granite can be found. If the Huronian formations are followed eastward along the strike near the southern edge of the ridge which they form, a number of places may be observed where granite intrudes the Ajibik quartzite. Finally, near the center-line of Section 25, which passes just west of the west end of Lake Palmer, there is a high ridge of Siamo slate extending eastward and westward along the northern side of the lake. The south side of this ridge, west of the lake, forms a steep cliff of nearly bare rock, and from the base of this cliff can be observed numerous dikes cutting the Siamo slate. Examined in detail, the slate is found to be highly granitized in places and traversed by many dikes and injections. Some of the dikes branch and finger out, while others pass into *lit-par-lit* injections and then again break across the stratification. The largest of the dikes, which shows very plainly along the face of the cliff, does not appear on the surface of the outcrop. This is a rather characteristic behavior of some of these dikes and domelike intrusions and was observed in a number of other places where exposures in section were obtainable.

Much of the Huronian material described in these portions of

Sections 25 and 26 was originally mapped as Palmer gneiss and designated Archean,[4] but later the true character of part of the Huronian formations was recognized.[5] However, the source of the granite dikes and injections was not accounted for at that time.

6. *The Palmer area.*—East and south of the area just described the Middle Huronian Ajibik quartzite and the Palmer gneiss, which has been designated Archean, are extensively intruded by granite which frequently appears as *lit-par-lit* injections, dikes, and domelike masses. These relations are well shown close to state road No. 35, which passes southeasterly through this area. Close to this road, about $\frac{1}{2}$ mile south of Palmer, the Ajibik quartzite, much metamorphosed, is exposed, particularly to the west of the road, between it and Werner Creek. In the latter locality the quartzite is intruded in a number of places by granite. Southeastward along the road numerous injections, dikes, and domelike masses of granite, intrusive into the Palmer gneiss, are exposed. The *lit-par-lit* injections have a fairly constant trend approximately N. 45° W. Farther northward from the road many granite intrusions in the Palmer gneiss are to be seen. However, within a distance of about $\frac{1}{4}$ mile, the Palmer gneiss ends and the Middle Huronian Ajibik quartzite, striking approximately N. 75° W., occurs instead. This quartzite also is intruded by the granite. On the surface of one outcrop an intrusion appears as a narrow dike, but on the side of the cliff, where it is exposed in section, it is seen to be a domelike mass. This relation is shown in Figure 4. North of this quartzite the iron-bearing formation is found, but it differs here from the typical Negaunee iron formation in that it contains coarse graywacke bands toward its lower part and generally becomes conglomeratic at its base. It is followed southward, however, by the Ajibik quartzite without any apparent change of strike between the two formations.

It will be recalled that these Middle Huronian formations strike about N. 75° W., but that the injections in the Palmer gneiss have a general trend approximately N. 45° W., and that both the Palmer gneiss and the Ajibik quartzite are intruded by the granite. This

[4] *Mono. U.S. Geol. Surv.*, Vol. XXVIII (1897), Atlas Sheet IV.

[5] C. R. van Hise and C. K. Leith, "The Geology of the Lake Superior Region," *ibid.*, Vol. LII (1911), pp. 255–56 and Pl. XVII.

relationship between the strike of the Ajibik quartzite and the trend of the injections in the Palmer gneiss very strongly suggests the unconformable relations that exist between the Middle and Lower Huronian formations on the northern limb of the Marquette synclinorium about 6 miles directly north of the area under consideration.[6] In view of the fact that some of the Middle Huronian formations were formerly designated Palmer gneiss, it may well develop that at least a part of the Palmer gneiss in this area near Palmer may represent Lower Huronian formations, possibly some of the

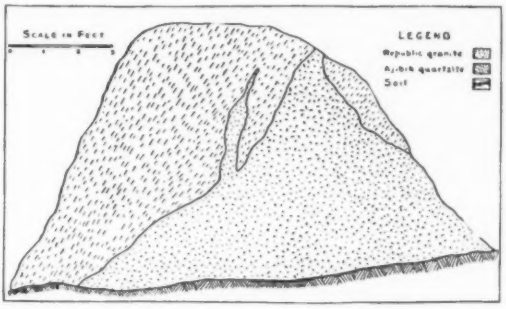

FIG. 4.—Sketch of section exposed near Palmer showing the Republic granite intruding the Ajibik quartzite from beneath (looking eastward).

lower slate and graywacke members of the Mesnard quartzite series. If this should prove to be the case, the erosion relations on the southern limb of the Marquette synclinorium would approximate those found on its northern limb.[7]

7. *The Greenwood area.*—In the vicinity of Greenwood, C. O. Swanson[8] cites records of drill holes indicating that the granite intrudes the Upper Huronian Goodrich quartzite. One of these drill records shows that granite was encountered after passing through about 200 feet of Upper Huronian formations, including part of the Goodrich quartzite, and that after passing through some 50 feet of granite the Goodrich quartzite was again encountered.

[6] *Ibid.*, Pl. XIX. [7] *Ibid.*

[8] *Report on the Portion of the Marquette Range Covered by the Michigan Geological Survey in 1929*, pp. 13–14 and Secs. II and VIII.

8. *The Beacon area.*—In this area, south of Beacon, what is essentially the contact between the granite and the Middle Huronian formations was traced for nearly a mile. The general relations are shown by the accompanying map (Fig. 5). This map was made by means of compass and counting paces, and absolute accuracy is not claimed. It will be observed that the general strike, as shown by the iron formation, is approximately N. 75° E. and that the granite apparently is cutting discordantly across the strike. It will be observed, also, that the Ajibik quartzite is engulfed and entirely sur-

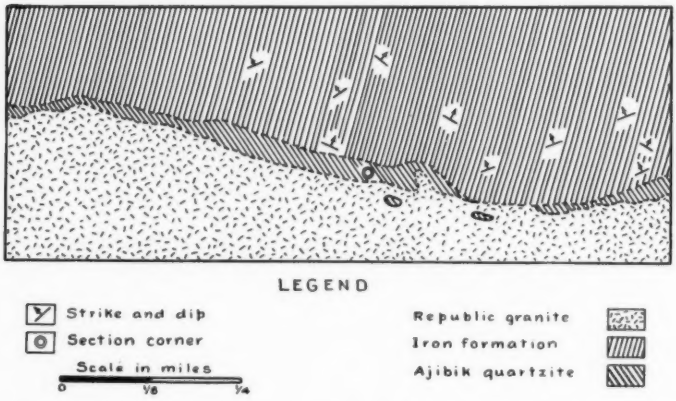

FIG. 5.—Map of an area near Beacon showing the contact between the Middle Huronian formations and the Republic granite. Note the quartzite engulfed by the granite.

rounded by granite in places. Much of the iron formation is strikingly metamorphosed into a grünerite-garnet-magnetite rock, whereas the Ajibik quartzite is highly altered and transformed into a gneiss in places. Microscopic examination of a series of specimens collected just west of this area shows a gradual transition from gneissoid granite to unmistakable quartzite.

9. *The Lake Michigamme area.*—The Upper Huronian formations, which crop out in the areas south of the east arm of the lake, south of the west arm of the lake, and on islands and along the shores in the southern part of the lake, frequently show evidence of having been intruded by granite and are often metamorphosed into staurolitic

and garnetiferous biotite schists containing a considerable amount of tourmaline and andalusite in places. In a belt extending through parts of Sections 2 and 3, east of the south arm of the lake, and but a few miles west of the Beacon area just described, the Upper Huronian formations have been metamorphosed into gneiss and have been intruded by granite in a number of places. These intrusions are in the form of dikes and small knoblike masses (Fig. 6). A considerable part of this gneissoid material, as shown by microscopic examination and field study, was originally a quartzite, probably the Goodrich quartzite. The formations above the Goodrich quartzite crop out along the shores of the lake, west of this location, and these formations are underlain eastward by a quartzite. This would be the normal relation for the Goodrich quartzite. Microscopic examination of specimens in this vicinity shows gradation from quartzite through gneiss into gneissoid granite.

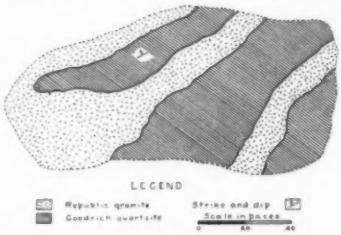

FIG. 6.—Sketch map of an outcrop near Lake Michigamme showing the Upper Huronian Goodrich quartzite intruded by the Republic granite.

On small islands in the southern part of Lake Michigamme small knoblike and dikelike masses of quartz containing some feldspar and muscovite, and frequently a considerable amount of apatite, occur in the Michigamme formation. Around these masses, practically at the contact between them and the Michigamme formation, andalusite crystals about 2 inches long and $\frac{1}{2}$ inch across were found, and much staurolite and garnet occur on the islands.

Along the south side of the western arm of the lake numerous dikelike intrusions, from a few inches to a foot or more in width, cut the Michigamme formation. Some of these consist almost entirely of feldspar; others consist essentially of quartz with small quantities of feldspar. They frequently contain crystals of apatite $\frac{1}{8}$ inch across. Microscopic examination of the sediments cut by these intrusions shows that they characteristically contain a considerable amount of tourmaline. Extending westward for about 2 miles there is a belt of staurolitic and garnetiferous Michigamme formation which, however, shows no dikelike intrusions in the outcrops.

In view of the relations mentioned at other places—namely, that the granite sometimes does not crop out at the surface or does occur as small dikes where it is known to exist as masses of considerable size beneath the surface—it appears likely that the granite extends, tonguelike, beneath this area west of Lake Michigamme. Certain structural relations of the granite, to be cited later, tend to substantiate this probability.

AGE OF THE GRANITE

From the foregoing descriptions it is apparent that the Lower, Middle, and Upper Huronian formations are intruded by granite. In the northern part of the area there is no horizon marker above the Upper Huronian that would serve as an index of the age of the granite. Some diabase dikes of the Keweenawan type cut the granite, but their age is not definitely known. In the southern part of the area, however, near Felch, Cambrian (Potsdam) sandstone is found resting on the granite and not intruded by it. It seems clear, therefore, that the granite is post–Upper Huronian and pre-Potsdam.

There is a possibility of several periods of intrusion, and throughout the entire course of the field work this was kept constantly in mind. No evidence was found, however, for more than one period of intrusion. On the contrary, the evidence regarding the structural relations of the granite points toward but one period, and throughout its entire extent, except near its margins where much of it is gray to white, the granite has a rather uniform appearance. In many respects this granite and its marginal facies apparently are similar to some of the Killarney granites.[9]

There is also a possibility that part of the area may contain some pre-Huronian granite. No granite that could be definitely assigned to the pre-Huronian was found, however, although the recognition of such a granite in an area which, since pre-Huronian time, has been closely folded and extensively invaded by another granite would be extremely difficult.

NAME

It is suggested that this post-Huronian granite, which has been known as a part of the Archean Southern Complex, be designated

[9] W. H. Collins, "North Shore of Lake Huron," *Can. Geol. Surv., Mem. 143* (1925), pp. 86–87.

the "Republic granite," as it is well exposed, easily accessible, and shown in most of its relations in the vicinity of Republic. Along state road No. 45 and in the adjoining fields there occur large masses of this granite which exhibit its characteristic porphyritic character. The pegmatitic facies of the granite may be seen in a number of places, and the fine-grained contact phase occurs at Republic although there it is red instead of gray. Also, as previously stated, the contact with the Middle Huronian may be found in this vicinity. Moreover, basic dikes cutting the granite may be observed.

STRUCTURAL RELATIONS

As the Republic granite is observed in the field, one of the things constantly brought to one's attention is the striking parallelism of the feldspar phenocrysts. Most of these have a general northwesterly trend. Although the determination of the primary structure of the granite is really a separate problem requiring many careful field measurements, some work of this nature was undertaken, using the method developed by Professor Hans Cloos, of the University of Breslau, Germany, and practiced in this country, especially by Robert Balk. Briefly this method consists of measuring and plotting mineral, *schlieren*, joint and dike orientations, and from these data drawing conclusions regarding the primary structure of the intrusive. For accurate work hundreds of observations are necessary. However, in the area under investigation, where there are no quarries in the granite and outcrops of the proper character are not readily found, the taking of so many observations would require a very great amount of time. Nevertheless, about one hundred observations were recorded and plotted. The results, as shown by Figure 7, indicate that the general strike of the mineral and *schlieren* parallelism is northwest, that these features mostly dip northeastward rather steeply, and that there are two principal joint directions, one northwest and the other northeast. The northwest joints are more abundant than the northeast ones and generally dip northeast the same as the feldspars. Although the few dikes observed have various trends, the majority trend northeast.

These structural relations, when interpreted in the manner developed by Professor Cloos, indicate a regional tectonic stress directed

southwestward. This stress would cause a stretching of the granite in a northwest-southeast direction, and a definite mineral orientation in that direction, and might account for the possible extension of the granite beneath the Michigamme formation, as previously suggested. Further, the northwest-trending joints would be compression joints during the earlier history of the intrusion and would

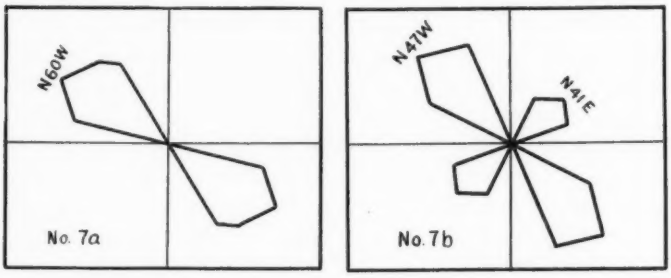

FIG. 7.—Diagrams showing the strike of feldspar phenocrysts and *schlieren* (No. 7*a*) and the strike of the joints (No. 7*b*) observed in the Republic granite.

not in general be dike-bearing joints, whereas the northeast-trending joints would be tension joints and would contain dikes. These conclusions are well in accord with conclusions reached by Robert Balk in a similar instance involving regional tectonic stress in which, however, the force was directed toward the northwest.[10] Further, they are in accord with the general pre-Cambrian structural relations of the part of the Lake Superior region involved and apparently, also, with the relations in Ontario.[11]

[10] "A Contribution to the Structural Relations of the Granite Intrusives of Bethel, Barre, and Woodbury, Vermont," *Vt., State Geol. 15th Rept.* (*1925–26*) (1927), p. 59, Fig. 11, and pp. 69–71.

[11] *Loc. cit.;* T. T. Quirke and W. H. Collins, "The Disappearance of the Huronian," *Can. Geol. Surv., Mem. 160* (1930).

AN INTERPRETATION OF BUNSEN'S GEYSER THEORY

WILLIAM H. SHERZER
Michigan State Normal College

ABSTRACT

The geyser theory of Bunsen appears to the writer to be imperfectly presented, as found in English and American textbooks on geology. Two erroneous assumptions are to be found. First, that the relief of pressure assumed in this theory is due to an overflow from the top of the water column or to expansion into a basin with resultant shortening of the column. Many splendidly functioning geysers have no visible overflow preceding an eruption and very many have no basins. Second, it is assumed that the mere lifting of the upper portion of the water column without overflow would relieve the pressure below sufficiently to bring on an eruption. As long as the head remains the same, obviously there could be no such relief of pressure. The observations of Bunsen and Descloizeaux upon the Great Geyser of Iceland established the existence, about halfway down in the tube, of what may be called the "critical zone." Here the actual and boiling temperatures come very close together as the conditions for an eruption are being slowly attained. Steam bubbles are first formed in the lowest depths of the system and when powerful enough slightly elevate a portion of the water at the critical zone, bringing it to a level where it can flash into steam, thus starting the eruption and relieving the pressure on the lower portion of the water column.

FIELD METHODS EMPLOYED

Bunsen and Descloizeaux found the Great Geyser to have a nearly circular basin, 16–18 meters in diameter, and 1.57 meters deep. In the center of this basin was a cylindrical well about 3 meters in diameter. Hooker had previously compared this basin to "a saucer with a circular hole in the middle." This circular hole was the opening to the geyser tube, or well, and appeared to be approximately the same diameter as far down as could be observed after a major eruption. About the margins of the basin occurred some small clefts through which a very small overflow from the filled basin took place during the "upheavings" between major eruptions. Such disturbances and small overflows of the water column took place about every hour and a half and led to no eruption (Descloizeaux). The central well, or real geyser tube, was found to have a depth of but 21.5–23.5 meters. The observations of these two men upon the Great Geyser covered ten major eruptions, separated by an average interval of twenty-eight hours and thirty-five minutes and ranging from ten hours and fifty-five minutes to forty-six hours and ten min-

utes. The eruptions lasted from five to seven minutes, and upon two occasions the jets reached heights of from 47 to 49 meters as determined by a theodolite. Five observations were made upon the temperatures of the main water column by attaching self-registering thermometers enclosed in metallic cases at approximately equal distances along a cord and lowering them into the tube for thirty minutes. The theory of Bunsen is deduced entirely from these thermometric studies. The elevation of the geyser group above Reykjavik was approximately 110 meters (barometric).

OBSERVATIONAL DATA ON THE GREAT GEYSER

The results of the observations by Bunsen and Descloizeaux are compactly presented in Table I.[1]

TABLE I

TEMPERATURES IN DEGREES C. HEIGHTS FROM BOTTOM IN METERS

First Experiment		Second Experiment		Third Experiment		Fourth Experiment		Fifth Experiment	
Temp.	Height	Temp.	Height	Temp.	Height	Temp.	Height	Temp.	Height
95.0	19.70	82.6	19.20	85.2	19.55	84.7	19.55	.....	.....
109.0	16.30	85.8	14.40	106.4	14.75	110.0	14.75	103.0	13.50
x	12.90	.....	.....	.....	.....	.....	.....	.....	.....
121.1	9.50	113.0	9.60	120.4	9.85	121.8	9.85	121.0	9.70
121.6	6.00	122.7	4.80	123.0	5.00	x	5.00	x	4.90
.....	.....	123.6	0.30	127.5	0.30	126.5	0.30	122.5	0.30
.....	Bottom	.....	Bottom	.....	Bottom	.....	Bottom	.....	Bottom
Mean temp. 112.68°		Mean temp. 102.30°		Mean temp. 108.33°		Mean temp. 109.19°		Mean temp. 108.83°	

From these data and a few additional observations mentioned below, certain generalizations appear to be justified.

1. In all cases the temperature of the water in the tube was found to increase with the depth, although not regularly. That of the basin was subject to considerable variation, ranging from 76.0° to 89.7°, depending upon the state of the atmosphere, force and direction of the wind, elapsed time since the last eruption, etc. The average temperature was assumed to be 85° C.

[1] M. A. Descloizeaux, "Observations physiques sur les principaux geysers d'Island," *Annales de chimie et de physique*, 3d ser., Vol. XIX (Paris, 1847), pp. 444–70.

2. At all depths, aside from a few minor disturbances, the temperature of the water in the tube gradually rose during the interval between eruptions.

3. Up to within a few minutes before an eruption the temperature of the water in the tube was not observed to reach the boiling point for that depth and atmospheric pressure.

4. Major eruptions were always preceded by premonitory rumblings and detonations which became more violent with the approach of the eruption. These were noted to begin about four or five hours after such major eruptions and were more pronounced at intervals of one to two hours, finally becoming extremely violent.

5. By throwing fragments of paper into the middle of the basin it was noted that there was an ascending column of heated water at the center which spread over the surface of the basin to its edges and then back along the bottom of the basin to the geyser tube, occasionally appearing "as if it were forcibly drawn back."

6. At a level about 10–11 meters down, mid-depth of the tube, there occurred what may be appropriately termed the "critical zone," since here the actual temperature and the boiling temperature for depth most nearly approached one another "in proportion to the approximation of the period of a great eruption." This fact is indicated graphically in Figure 1 in which the smooth curve marked "1," leading from the right upward to the left represents the actual boiling temperatures for depth. The broken lines "2," "3," and "4," plotted from left to right and downward, are based upon the data obtained in the fourth, third, and second "experiments," respectively. These graphs will be found to be of the greatest significance in elucidation of the theory.

7. Rock fragments suspended in the tube at various depths and marked for identification were ejected at the time of the eruption if placed above the critical zone, but not if suspended in the lower half of the tube. One thermometer so placed survived an eruption and registered a temperature of 124.24° C. at a height of 4 meters above the bottom of the tube.

"On considering all the phenomena presented by the eruption of the geysers, we cannot for a moment doubt that the main seat of the mechanical force, by which the mass of water is thrown up and con-

verted into boiling foam, is actually situated in the funnel of the geyser" (Bunsen).

8. Another phenomenon not noted by Bunsen in connection with the Great Geyser has a very intimate relation to the theory under discussion. When certain geysers are approaching the conditions necessary for a major eruption, premature action may often be induced by throwing into the tube rock fragments, or soil, or by violently

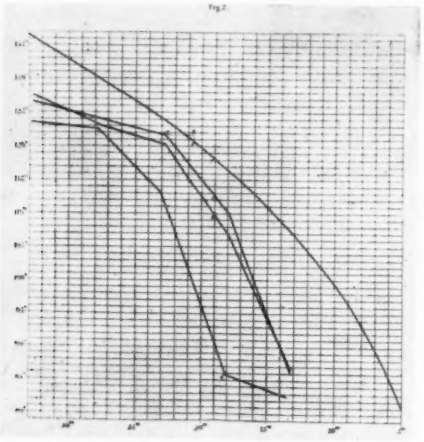

FIG. 1.—Bunsen's original graphs showing relation of actual temperatures of water in the geyser tube to boiling temperatures computed for pressure (graph No. 1). Curve No. 2 shows observed temperatures twelve minutes before an eruption; No. 3, four hours and fifteen minutes before; and No. 4 twenty-two hours and fifty minutes before the next major eruption. Critical zone level indicated at *ab*.

agitating the water in the basin or upper portion of the tube. The Great Geyser is said to refuse to respond to any such attempts to produce a premature eruption, but the neighboring Strokkur with a totally different type of tube responds in from twenty to thirty minutes, and this may be successfully repeated several times a day. In the case of the artificial geyser model it may be noted that a small amount of shaking or disturbance of the tube often will bring about the same result.

FUNDAMENTAL ASSUMPTIONS

The ability of the theory under consideration to explain satisfactorily the phenomena of geysers rests upon certain assumptions. Possibly some of these do not have universal application without preventing the operation of the geyser mechanism. They probably, however, all apply in the case of the Great Geyser of Iceland.

1. The necessary water for the operation of the geyser is atmospheric.

2. The heat necessary for the action is derived from beds of buried lava which are thus very slowly being cooled by conduction to such percolating water and to the adjacent cooler beds.

3. As far as could be observed after major eruptions and from soundings, the geyser tube was practically cylindrical throughout, with no indications of lateral expansions, caverns, or tunnels, into which the rising column of water might overflow and escape.

4. Leading out from the bottom of the geyser tube, irregularly and indefinitely, in the heated water-bearing strata, channels and fissures are assumed. Through these the water is periodically supplied under hydraulic pressure. The range in the time interval of the major eruptions seems due to the irregular supply of replenishing water rather than to variations in the temperature of the heated beds.

5. Owing to the small diameter of the geyser tube in relation to its length, the convection currents are seriously impeded and the column of water is irregularly heated in consequence, giving rise to the critical zone above noted.

6. Although the actual temperature and the boiling temperature approach each other very closely at the critical zone, they are believed to meet first in the lower depths ordinarily beyond reach of observations. Here, then, is steam first produced, the bubbles of which rise into the relatively cooler water where they collapse and give rise to the rumbling and detonating that precedes an eruption. When these bubbles are large enough or abundant enough they force the water up into the basin possibly causing some overflow, but no eruption, and with their collapse cause the water to be drawn back, more or less forcibly, into the tube. From the latent heat of this

steam assisted by convection the water higher in the geyser tube is heated from below and thus the stage is set for the eruption.

7. Any theory that can satisfactorily explain geyser phenomena must account for the sudden, almost instantaneous, creation of steam somewhere along the tube.

8. In the theory under consideration it is assumed that there is no overflow, either from the basin or underground, immediately preceding the eruption. This is also probably true of the great majority of the geysers in Yellowstone National Park.

9. Simply lifting the upper half of the column of water in the tube without overflow into a basin, or over its rim, could not appreciably relieve the pressure upon the lower half of the column. This would be true regardless of how much steam or other gas had been mixed with the water during the simmering process preceding an eruption.

10. If the theory of Bunsen is to be used in general in explaining geyser activity, it is necessary to assume the existence of a critical zone in every active geyser for which no explanation is as yet available. Here is a promising field for some student of thermodynamics. In the case of the artificial model this zone is imitated by placing a heated collar somewhere near the center of the tube.

11. Finally, it is assumed that, with the formation of steam adjacent to the heated lava at the level of the critical zone, a further production of steam takes place which ejects the upper portion of the column as superheated water, relieves the pressure, and permits the remaining water to flash into steam and with its escape to bring the eruption to a close.

INTERPRETATION OF THEORY

In the original presentation of his theory Bunsen failed to place the emphasis sufficiently upon the importance of the critical zone, evidently leaving this for the reader to gather from the data presented. To this omission may be attributed the very general misunderstanding that has been so long perpetuated in most geological texts. From the preceding array of observations and assumptions it remains to piece together the happenings in the tube that can satisfy the physical conditions essential to an eruption. Since the source of the energy for the activity as well as the water itself lies first in the extreme

depths of the tube and far beyond the limits of observation, we may well quote at some length from Descloizeaux.[2]

Suppose that the column of water in the central basin communicates, by a long and sinuous channel, with the space, be it what it may, which receives the direct action of the subterranean heat: after an eruption during which a projection of a great quantity of water and vapour has taken place, the lower parts of the liquid mass are cooled, and the steam which is formed in the reservoir submitted to the action of the heat has a less tension than that at which the weight of the central column and that of the atmosphere are in equilibrium; this vapour, as it forms, becomes condensed in contact with the water which fills the sinuous channel, and it imparts to that water its latent heat. The increase of temperature of the water of the channel is transmitted by degrees to the lower part of the central column where the thermometer can reach; but this increase is retarded by the atmospheric air and the other gases which accompany the vapour; however, at the lapse of a shorter or longer time, the water of the channel must boil, and the steam which continues to form can not longer condense there.

From this point Bunsen is quoted:[3]

The greater portion of the Icelandic thermal springs exhibit the peculiar but easily explained property of giving rise periodically, at certain points in the water of the thermal basins, to a number of large bubbles of vapor, which become suddenly condensed on rising into an upper and cooler stratum. This invariably occasions a slight detonation, accompanied by a hemispherical elevation, and an instantaneously succeeding depression of the surface of the water. Even the Great Geyser is characterized by a periodic succession of the detonations of vapour, beginning about four or five hours after a great eruption, and continuing, at intervals of from one to two hours, until the next eruption; immediately preceding which they occur in rapid succession and extreme violence. . . . . The periodic elevation of the masses of water effected in the geyser by these means appears rarely to exceed 1 or 2 meters, if we may judge by the average height of the mass of water which is forced from the mouth of the tube in the form of a conical elevation.

In the writer's interpretation these bubbles of steam eventually become large enough and powerful enough to lift a *portion* of the water column, either at one side of the tube or possibly at the center, to a height above the critical zone at which its actual temperature

[2] "Physical and Geological Observations on the Principal Geysers of Iceland," *Philosoph. Mag.*, 3d ser., Vol. XXX (London, 1847), p. 401.

[3] R. Bunsen, "Über den innern Zusammenhang der pseudovulkanischen Erscheinungen Islands," *Annalen der Chemie und Pharmacie*, Vol. LXII (Heidelberg, 1847), pp. 1–59; complete translation of foregoing by Dr. G. E. Day, *Chem. Repts. and Mem.*, ed. Thomas Graham (London, 1848), pp. 349–50. Pl. II.

somewhat exceeds that of the boiling temperature for this new depth. Of course, a similar volume of water lying in this critical zone would drop to a lower level and replace the water so elevated. This may be seen best by examining Figure 2, copied from Tyndall,[4] in which this zone is indicated as lying between the double parallel lines *A* and *B*. The observed temperatures here represented, with corresponding depths and boiling temperatures, are those given by Bunsen and

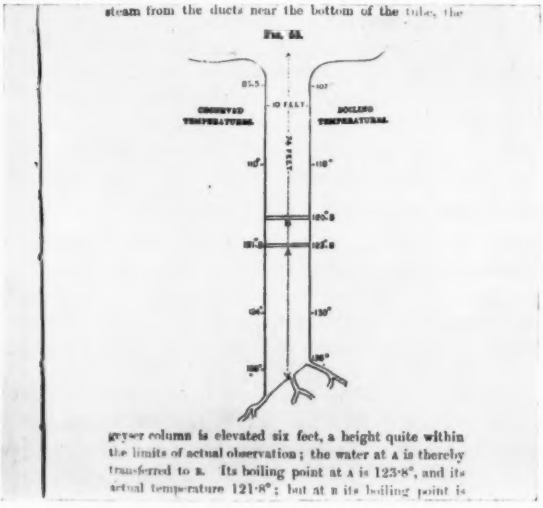
steam from the ducts near the bottom of the tube, the
FIG. 53.
85.5
107
10 FEET.
OBSERVED TEMPERATURES.
BOILING TEMPERATURES.
24 FEET.
119°
130°
geyser column is elevated six feet, a height quite within
the limits of actual observation; the water at A is thereby
transferred to B. Its boiling point at A is 123·8°, and its
actual temperature 121·8°; but at B its boiling point is

FIG. 2.—Tyndall's diagram to explain Bunsen's theory of geyser action as determined from studies upon the Great Geyser of Iceland. The critical zone is indicated as lying between *A* and *B*, and the observed temperatures are those obtained in the "Fourth Experiment," twelve minutes before an eruption.

Descloizeaux in their "fourth experiment," twelve minutes before a major eruption, the second (from the bottom) temperature reading being interpolated from the graph shown in Figure 1. As the time approaches for the eruption, the actual temperature and the boiling temperature more nearly approach each other, the column of water in the critical zone becomes shorter, bringing *A* and *B* still nearer to each other, necessitating by that amount less upward movement of a

[4] *Heat a Mode of Motion*, 6th ed. (New York, 1905), p. 169.

portion of the water from level *A* to level *B*. When this water at temperature 121.8° finds itself at a depth requiring but 120.8° for boiling, it flashes into steam, sending the preliminary jets of water into the air, relieving the pressure upon the entire column and inaugurating the real eruption. Much of the water below temperature 100° is ejected as such, probably lying entirely above the critical zone, while that below with its greatly relieved pressure passes into and escapes as steam, entirely emptying the geyser tube and its branching channels. Could we but measure the amount of water and steam ejected at the time of a major eruption we would have a clue to the contents and size of the system.

As the stage was being gradually set for an eruption in the region of the critical zone, it can be readily imagined that stones dropped into the geyser tube might disturb the water there sufficiently to give the same effect as the rising steam bubbles and thus hasten an eruption. Several points should be here emphasized:

1. The theory calls for no overflow from the edge of the basin, and if any occurs in certain cases it is but incidental. It is quite possible that such overflow and consequent relief of pressure might, under certain circumstances, cause a boiling spring to erupt, but this is not a part of Bunsen's theory.

2. The relief of pressure necessary to start an eruption is not due to the slight shortening of the column that might result from the partial filling of the basin. This would probably be fully compensated for by the conical or hemispherical swelling that usually just precedes an eruption. Then, too, many geysers have no basin and do not visibly overflow.

3. The mere lifting of that portion of the water column lying above the critical zone, without any overflow or expansion into a basin, cannot appreciably relieve the pressure upon the remaining water of the column, and this is true no matter how much steam or other gas it may contain. Two assumptions, at least, may be considered:

a) As the water column is lifted by steam pressure below, the critical zone might be raised so as to remain the same distance from the surface of the water in the tube. This would elevate any given stratum of superheated water but would also call for a corresponding

elevation of the boiling point, and the two temperatures would still be as far apart as before, the head remaining the same.

b) Should the critical zone remain at the same level in the tube and the water from the next lower position with actual temperature 124° be lifted to *A* (Fig. 2), where the boiling temperature *was* 123.8°, no steam would be produced, since the head now calls for a boiling temperature of 130°.

With this interpretation in mind the writer has scrutinized the explanations of Bunsen and Descloizeaux to see if they contain anything that might justify such assumptions relative to the functioning of the critical zone. Possibly we have in the following two quotations the real crux of the theory.

"A glance at the graphic representation given in fig. 2 [here Fig. 1] will, however, show that such an upheaval as this is insufficient to raise any stratum of water to an elevation where it could be brought to a state of ebullition (in consequence of the diminution of pressure existing there), until the mass of water would, by gradual heating, be brought to the temperature of the broken line, 2; and such is actually observed to be the case a few minutes before an eruption. All the other upheavals preceding this period would only be able to drive partially the lower heated masses of water by a sudden impulse into the upper part of the tube of the geyser, where these masses are brought into a state of ebullition, owing to the diminution of pressure."[5]

"If the vapour formed in the subterranean reservoir is not condensed entirely at its contact with the water of the sinuous channel of which I have spoken, and if it is accompanied with a sufficient quantity of atmospheric air, or of other gases, it must in the end divide the liquid column, and escape at the surface, producing those subterranean detonations and those jets which all travellers have remarked."[6]

EXPERIMENTS WITH GEYSER MODELS

Probably most geological laboratories have one or more models that appear to illustrate artificially the principles of Bunsen's theory, and certainly they do erupt. The writer has one consisting of a cylindrical tin tube, 1.85 meters long and approximately 33 millimeters in diameter, which upon one occasion sent a jet to a ceiling some 6 meters above the mouth of the tube, or slightly over three times its own length. This is some 50 per cent better than was the

[5] Bunsen, (trans. G. E. Day), *op. cit.*, p. 350.

[6] Descloizeaux, "Physical and Geological Observations on the Principal Geysers of Iceland," *op. cit.*, p. 402.

best-observed effort of the Great Geyser. Tyndall credits Dr. Bromeis of Marburg with having constructed the first artificial geyser, but gives no date nor reference. Since he was a pupil of Bunsen and studied at Marburg it seems very probable that he is correct. However, J. Müller of Freiburg,[7] as early as January, 1850, described his apparatus with thorough approval of Bunsen's theory. Müller's geyser model had a length of 1.5 meters and a diameter of 12 centimeters, was heated at the bottom and about one-third up with glowing charcoal, and was able to project the water periodically to a height of 1–2 feet. Had a nozzle been inserted so as to contract the mouth opening, a much greater height could have been secured. However, tubes with the larger diameters are more sluggish in their action, because of the greater quantity of water to be heated and because the convection currents are less impeded. The writer has a second model with a length of 1.875 meters, diameter 9 centimeters, heated by two circles of gas jets, one 16 centimeters and the other 87 centimeters from the base. If the gas jets are rather short, just reaching the tube, the water in the apparatus will simply simmer, but if the lower is turned on strongly a vigorous burst of hot water will be ejected in from 30 to 40 minutes to a height of 2–3 feet. Some 3–4 liters of water are needed to replace the loss, and when this is added and a nozzle inserted, the eruptions follow at the rate of about one a minute, the jets reaching a height of some 6 feet. In the case of the smaller bore tube mentioned above, the primary eruption occurs in seven to eight minutes, and with the nozzle in place, the jets are higher and successive eruptions more frequent as more and more water is lost. Without the nozzle an eruption of the Excelsior Geyser type is obtained while with it the Old Faithful and Bee Hive eruptions are simulated.

The writer has observed, as many others may have done, that eruptions may be secured with such apparatus by using a single burner near the base. The second upper burner has been supposed necessary in order to establish the critical zone in the water column. To get an eruption with but a single burner raises the rather disquieting question as to whether such models do really illustrate Bun-

[7] J. Müller, "Über Bunsen's Geysertheorie," *Annalen der Physik und Chemie*, Taf. 3, Fig. 6, Vol. LXXIX (Leipzig, 1850), pp. 350–53.

sen's theory, or whether they operate because of other thermodynamic principles. The writer inclines to the belief that when two burners are used the heat may be so adjusted that steam is first formed at or near the base of the water column and a critical zone established artificially by means of the second burner, the eruption being due now to true Bunsen principles. When but a single burner is used and an eruption secured, we must conclude either that a critical zone does actually exist in our model and hence the second burner is unnecessary or that the eruption calls for another explanation, such as overflow with relief of pressure or, more plausibly, superheating of the water. It was discovered many years ago by Donny that when water is freed from its dissolved air, or other gases, as by boiling, it may be heated some 30°–40° C. above its boiling temperature without the formation of steam, and then it finally lets go with explosive violence.

As far as the writer is aware no one knows just what takes place within the geyser tube. To get such data a pyrex tube could be utilized, and the water could be heated by electric units carrying a string of chemical thermometers. In this way the temperatures could be read and the formation of steam directly observed.

THE GEOLOGY ALONG THE PERENE AND TAMBO RIVERS OF EASTERN PERU

PHILIP W. CHASE
Milton, Massachusetts

ABSTRACT

A cursory survey of the geology along the Perené and Tambo rivers in Eastern Peru discloses interesting facts concerning the stratigraphy, lithology, and structure of the little-known eastern slopes of the Andean Cordillera.

Two conspicuous breaks parallel the mountain structure, the first occurring near San Ramon, where the mountains drop from a general elevation of 10,000 feet, to an elevation of 5,000 feet; the second occurring at the front range of the Cordillera, where the mountains end very abruptly along a north-south scarp and descend from 5,000-foot elevations. Evidence indicates that the first of these breaks is due to a normal fault having a vertical displacement of 3,000–5,000 feet; the second is due to difference in resistance to erosion betweeen the Tambo limestone and the Cascada red shales. This means that the younger sedimentary beds extend essentially unbroken and with only the gentlest folding from the San Ramon fault to and under the Amazon lowlands.

The geologic column includes formations of volcanic breccias and conglomerates, sandstones, limestones, and red shales, ranging in age from Carboniferous (?) to Quaternary. Intrusions of coarse-grained granite occur along the line of structural weakness marked by the San Ramon fault. The formations have been correlated with those of the Central Peruvian Andes and with those of parts of Eastern Peru. Striking similarity is noted in the lithology and stratigraphy of these regions, suggesting that the Andean geosyncline extended as far east as the Ucayali River and as far north as the Marañon.

INTRODUCTION

The geology along the course of the Perene and Tambo rivers in Eastern Peru is of general interest, not only because the district may be of future importance as a producer of petroleum, but also because it discloses the structure of the eastern Andean slopes, an area that has received little detailed geological study. With the exception of Bowman's *The Andes of Southern Perú*[1] and Singewald's "The Geology of the Pichis and Pachitea Rivers"[2] and "Pongo de Manseriche,"[3] there has been no work published on the geology of the Peruvian montaña, as this section of the Andes is termed. Bowman's work was done along the 73d meridian, some 150 kilometers southeast of the area treated in the present paper. The Pongo de Manseriche on the Marañon River, the geology of which is described by

[1] I. Bowman, *The Andes of Southern Perú* (New York, 1922 (?)).

[2] J. T. Singewald, *Bull. Geol. Soc. Amer.* Vol. XXXIX (1928), pp. 447–64.

[3] *Ibid.*, Vol. XXXVIII (1927), pp. 479–92.

Singewald, lies approximately on the 5th parallel of south latitude, or 700 kilometers north of the Perene River. The Pichis and Pachitea rivers are situated in the same general district as the Perene and Tambo, but they flow at a small angle to the Andean mountain structure rather than across it.

It is the purpose of this paper to outline in a very general way the geology of an area situated midway between the districts studied by Bowman and Singewald, an area that cuts across the Andean mountain structure at right angles, and therefore provides a satisfactory cross section. The observations were made in the autumn of 1930, in the course of a journey made by J. D. Simpson and the writer from the headwaters of the Amazon River in the high Cordillera Occidental of the Andes to the port of Para in Brazil. There are no accurate maps of the area, most of which is still unexplored. Those by the Peruvian geographer Raymondi proved to be the most reliable and were used as a base. Corrections and additions were made wherever possible. In spite of the fact that the surface is covered with soil and thick jungle growth, rock outcrops are numerous along the river banks during the low-water stage, and in the cliffs which parallel the river.

The area considered lies along the course of the Perene and Tambo rivers from the town of San Ramon in the Chanchamayo Valley to the settlement of Atalaya on the Ucayali River, a distance of 180 kilometers. On the west this area is bounded by the Cordillera Oriental which rises abruptly above San Ramon to culminate in snow-covered peaks east of Jauja. The geology and physiography of the high country has been described in some detail by McLaughlin.[4] To the east the area is bounded by the Ucayali River and the flat-lying plain of the Amazon basin, where the thick mantle rock and jungle growth prevent the determination of the geology with any degree of accuracy.

The Perene and Tambo rivers flow almost due east across the area described at about the 11th parallel of south latitude. They are parts of a single river system, the Perene becoming known as the Tambo below the point where the Ene enters from the south. Farther east

[4] "Geology and Physiography of the Peruvian Cordillera," *Bull Geol. Soc. Amer.*, Vol. XXXV (1924), pp. 591–632.

the Tambo turns abruptly northward and, after being joined by the Urubamba which also enters from the south, is known as the Ucayali, which in turn is one of the main headwaters of the Amazon.

TOPOGRAPHY

The topography between San Ramon and the Ucayali is extremely rugged, with high relief. Erosion has been rapid on account of the heavy rainfall and the steep gradients. The general course of the

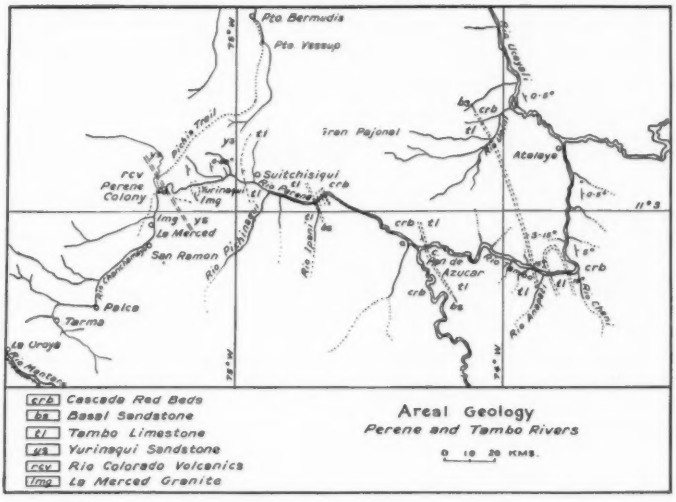

FIG. 1

Perene and Tambo rivers appears to be consequent upon the Andean uplift. Between the summit of the Cordillera Oriental and the Ucayali River the rivers drop 16,000 feet in a horizontal distance of 180 kilometers, so that the Perene-Tambo system was able to maintain its easterly course in spite of the discordant mountain structure which here runs northwest and southeast. The resistant formations rise as mountain ranges across which the river has cut narrow gorges, some of which have a depth of 2,000–3,000 feet. Farther east where the gradient is less, the river was unable to maintain its cross-cutting course, and at the elbow of the Tambo River the stream takes advan-

tage of the non-resistant red shales of the Cascada formation to change its course abruptly to the north, parallel to the structure. The tributaries of the Perene and Tambo, however, had neither the

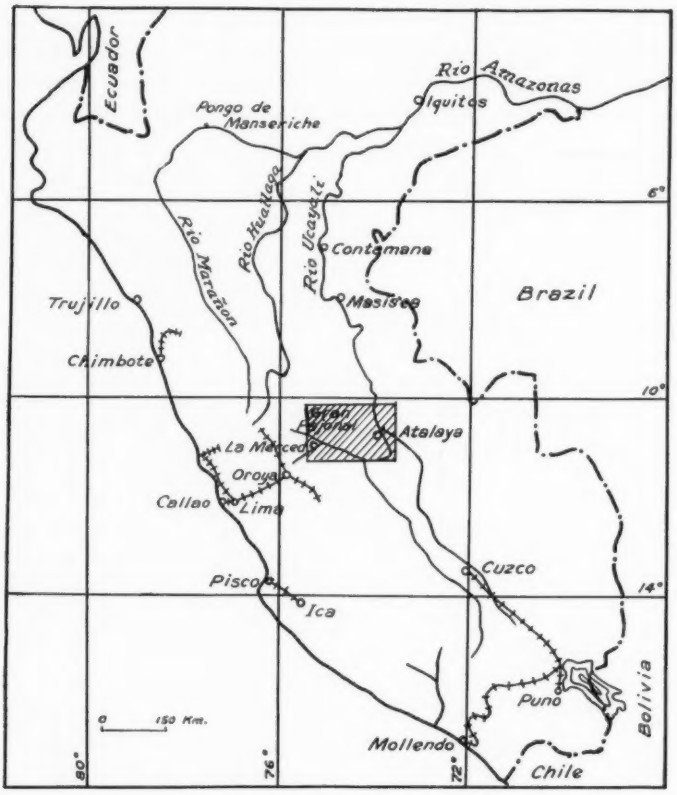

FIG. 2

necessary volume of water nor the gradient to follow a cross-cutting easterly course, and consequently have cut their valleys parallel to the structure in the softer formations. They all have a general north-south course.

There is evidence of an extensive peneplain found at an elevation of 3,500 feet and conspicuously developed to the north of the Perene River in the area known as the Gran Pajonal.

The eastern slope of the Andes shows two distinct breaks: one is at San Ramon, where the mountains drop from a general elevation of 10,000 feet to an elevation of 5,000 feet; the other is at the front range of the Cordillera, 120 kilometers farther east, where the mountains end very abruptly along a marked north-south scarp, and descend from 5,000-foot elevations to the featureless plains of the Amazon Basin (elevation 700 feet) within a distance of 17 kilometers.

The first break is believed to be due to an east-dipping normal fault. The fault contact itself is not visible, since the La Merced granite occupies the area through which the fault should pass. From the evidence at hand it appears probable that this faulting is of a late age (post-Puna of McLaughlin), possibly a renewal of movement along a pre-Tertiary zone of structural weakness which served as a locus for extensive igneous intrusions in Tertiary (?) times. The occurrence of Carboniferous (?) formations to the west of the fault zone in contact with Lower Cretaceous formations to the east indicates that the eastern side of the fault moved downward relative to the western side. The vertical displacement was at least 3,000 feet, equivalent to the thickness of the Jurassic limestone that is known to lie between the Lower Cretaceous and Carboniferous formations. If the general elevation of the summits east and west of the fault zone represent the remnants of a single peneplain (the Puna surface described by McLaughlin),[5] an uplift of almost 5,000 feet is suggested.

The second scarp is thought to be due to the difference in resistance of the Tambo limestone and the Cascada red shales. Bowman[6] mentions a strong front scarp fault to the south, which at first thought might be correlated to the break in question. However, it seems more probable that this fault continues its westerly course, and passes some distance to the south of the Perene River.

[5] D. H. McLaughlin, *Bull. Geol. Soc. Amer.*, Vol. XXXV, pp. 623–24.

[6] *Op. cit.*, p. 241.

GEOLOGY

The various types of rocks occurring in the area will be listed and described in the order of their occurrence, and an attempt will be made to correlate them with the formations recognized in adjacent regions by McLaughlin[7] and Singewald.[8] Fossils were collected, but the complete set was lost when one of the canoes capsized. The ages assigned must therefore be considered merely as an approximation, based on a comparison of the Perene River formations with those of the nearby Central Peruvian Cordillera with which the writer is

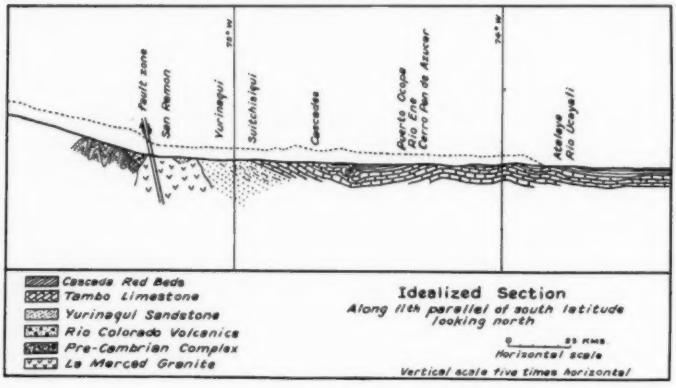

FIG. 3

familiar, as well as with descriptions of the formations of the Marañon and Bajo Ucayali rivers. The correlations given below are believed to be essentially correct, in spite of the lack of fossil determinations.

La Merced granite.—The Chanchamayo Valley, in which are located the towns of San Ramon, La Merced and Peruvian Colony, is cut in a stock of easily weathered, coarse-grained granite which contains conspicuous phenocrysts of quartz and pink feldspar. Some of the latter measure as much as 1 inch in length. The age of intrusion is difficult to determine. Steinmann[9] notes that in Peru nearly all of the igneous intrusions of post-Cambrian age are more basic

[7] *Op. cit.*, pp. 597–618. [8] *Op. cit.*, Vol. XXXIX, pp. 452–64.

[9] G. Steinmann, *Geologie von Perú* (Heidelberg, 1930).

than the earlier ones which are predominantly acidic. This generalization is probably true in the Cordillera, but it seems unwise to extend its use to the montaña until this area has been studied in greater detail. The La Merced granite appears to be intrusive into the Carboniferous and Comanchian sediments rather than to be surrounded by them as a granite monadnock. Whether it is associated with the deformation of the Cascada Red Beds, or whether it is independent of these diastrophic movements, is not obvious, but in either case it is probable that the intrusion took place in Tertiary times, at about the same period as the extensive intrusion of granodiorite and quartz monzonite in the Cordillera.

Rio Colorado volcanics.—Lying to the west of the La Merced granite stock is a series of purplish volcanic breccias, conglomerates, and sandstones. The contact with the granite is visible at the Rio Colorado, some 200 meters above the automobile bridge, and again at the Paucartambo River near the Peruvian Colony bridge. This formation closely resembles the Yauli volcanics of the sierra, tentatively assigned to the Carboniferous by McLaughlin,[10] and is believed to be the same formation.

Yurinaqui sandstone.—East of the granite stock, the Perene River enters a deep canyon cut in a thick series of sandstones, lavas, and limestones which are here termed the "Yurinaqui sandstone." Coarse-grained sandstone, strikingly cross-bedded and carrying conspicuous pebble layers, constitutes the bulk of the observed lower portion of this formation. Interbedded with the sandstones are thin layers of red shale and thicker flows of a dark red vesicular lava. The dominant rock weathers to a reddish brown. The upper part of this same formation is characterized by slightly finer-grained, light-colored sandstones interbedded with thin layers of limestone and red shale. Cross-bedding and pebble layers are common here as well as in the lower portion. The beds have a general easterly dip of about 20° although they show many gentle rolls and even occasional reversal of dips.

With more detailed study the Yurinaqui sandstone could undoubtedly be subdivided. An upper member, formed predominantly of light-colored sandstones with interbedded limestone, can be dis-

[10] *Op. cit.*, pp. 600–601.

tinguished easily from a lower series typified by cross-bedded, reddish sandstone, and vesicular lava flows. It seems probable that the Yurinaqui formation is from 5,000 to 10,000 feet thick, but any estimation of its thickness is uncertain on account of the varying dips and because of its termination against the La Merced granite intrusive.

Fossils were collected from the limestone members, but no determinations were made. In appearance the Yurinaqui sandstone in all its variations strongly resembles the Lower Cretaceous sandstone-basalt formation of the high Andes, named the Goyllarisquisga sandstone by McLaughlin.[11] Hot springs and tar seeps are found at various localities in the Yurinaqui sandstone. The correlated Goyllarisquisga sandstone is the chief coal-bearing formation of the sierra. If this correlation is correct, the formation shows a great increase in thickness as it is followed east, for McLaughlin[12] determined its thickness by actual measurement to be only 2,600 feet near the coal mines of Goyllarisquisga. Since in the Cordillera the upper contact is conformable with the overlying beds, it appears that these sandstones must occur as a thick lens, the center of which is located in the montaña, only the thin edge being found in the sierra. This fact may throw some light on the position of the geosyncline in Comanchian times.

Twelve kilometers above Yurinaqui the sandstone is intruded by a small stock of granite, identical with the main mass of the La Merced granite, and undoubtedly an outlier of it. This small plug is capped by sandstone, its roof being located about 1,500 feet above the river.

Tambo limestone.—Near the Indian village of Suitchisiqui the easterly dipping Yurinaqui sandstone is conformably overlain by the Tambo limestone. This formation is a shaly limestone, or a very calcareous shale carrying abundant dark, cherty nodules. It is typically thin-bedded and breaks into slabs 10–25 inches thick along the bedding planes which are markedly uneven and nodular. The shale layers are abundant and somewhat darker in color than the limestone members. Fossils are plentiful.

[11] *Ibid.*, pp. 605–8.

[12] *Ibid.*, pp. 605–6.

The Tambo limestone with an easterly dip of 5°–15° outcrops along the river from above Suitchisiqui to just below the Ipani River. Its thickness is at least 2,000 feet. It has been correlated with the Cretaceous Machay limestone of the sierra with which it appears to be identical in physical characteristics, mode of weathering, observed fossils, and sequence upon the Lower Cretaceous sandstones. It also closely resembles the descriptions given by Singewald[13] for the "Shale-Limestone" series of Pongo de Manseriche. Its age is late Lower Cretaceous probably overlapping slightly into the Upper Cretaceous.

Cascada Red Beds.—Below the Ipani River the Tambo limestone passes beneath the Cascada Red Beds, of which the Basal sandstone forms the lowest member. The contact appears to be conformable, although in the Cordillera there is known to be a marked unconformity between similar deposits and the older formations.

The Cascada Red Beds consist of a thick series of purple-red shales with interbedded limestone and sandstone. The relative proportions of shale to limestone-sandstone vary considerably, the shale comprising from 90 per cent to 20 per cent of the total, 85 per cent being the average. Where the red shales predominate, the individual limestone-sandstone beds are usually not over 3 feet in thickness, but are very conspicuous because they form rapids in the river. Where the limestone-sandstones predominate they tend to be somewhat thicker bedded, with the red shale intercalated in thin seams of from 1 to 6 inches. Parting along these shale layers is common, producing many dip slopes, which form the easiest way of differentiating the limestones of the Cascada formation from those of the Tambo formation.

At the base of the Cascada Red Beds is a thin formation of dense brown sandstone including minor conglomerate layers. The thickness is not over 500 feet, but its resistance to erosion and its distinctive color make it a useful horizon-marker. This very resistant sandstone overlying the likewise resistant Tambo limestone and directly underlying the soft red shales has controlled erosion and the drainage pattern to a marked degree. It is believed to be responsible for the cascades of the Perene River, the subsequent drainage of the Ene River Valley, the high mountain ranges just east of the Ene

[13] *Op. cit.*, Vol. XXXVIII, p. 487.

River, and also the steep eastern scarp of the Andean front range which lies just west of the Ucayali River.

For a distance of 17 kilometers below the Ipani River, the Red Beds are closely crumpled and folded; one of the anticlines brings the Tambo limestone to the surface for a short distance. The intense folding which produces vertical dips was accompanied by thrust faulting on a small scale. To the east the dips flatten out to form a shallow syncline.

The Cascada Red Beds have been assigned to the early Tertiary and Upper Cretaceous because of their position conformably overlying the Tambo limestone, and because of their similarity to the Casapalca Red Beds of the high Andes. No fossils were observed, nor are fossils common in the Tertiary formations of the Cordillera.

At the mouth of the Ene River is a steep cliff known locally as the Cerro Pan de Azucar, formed by the outcropping of the Basal sandstone and the Tambo limestone which dip 30° west at this point, and strike slightly west of north. From here to a point near the mouth of the Anapati River, a distance of 45 kilometers, the Tambo limestone outcrops along the river, forming almost vertical cliffs that rise to a height of 2,000 feet. The western limb of this anticline has a dip of 30° west, the eastern limb dips 5°–15° east, and there are at least three rolls or minor anticlines along its crest. No faulting or sharp folding was observed, all slopes being 20° or less.

Just above the mouth of the Anapati River the limestone disappears under the Tertiary Red Beds, only to emerge again in a shallow anticline 5 kilometers farther east. At the mouth of the Cheni River the limestone again dips below the Cascada Red Beds which here have a uniformly easterly dip of 10°. At this point the limestone-sandstone beds predominate and comprise 80 per cent of the formation, being distinguished only with difficulty from the underlying Tambo limestone. The presence of red shale partings is the only means of differentiation. Here, therefore, the Upper Cretaceous and the Tertiary formations not only are conformable but appear to be gradational. A short distance to the east, the normal Red Beds are found again, and continue north along the Tambo River all the way to its junction with the Urubamba. There is, however, a notable increase in brown sandstone as the river is followed

north. These sandstones are strikingly cross-bedded, and probably conform to the "Brown Beds" mentioned by Singewald[14] on the Pachitea River. Here they are interbedded with the red shales of the Cascada formation, and are therefore included under this more general heading.

Between Atalaya and the mouth of the Unini River outcrops are rare along the Ucayali River, the banks being formed of semi-consolidated clays, sands, and gravels. The few outcrops observed show the typical red shales, limestones, and sandstones of the Cascada formation, striking slightly west of north and dipping from 0°–5° east. North of the Unini River, with the exception of a single outcrop at the "Pozo" near the town of Pacaya (easterly dipping Red Beds), no outcrops of consolidated rock are seen until Contamana is reached 350 kilometers farther north. The prominent red banks seen at places along the river, notably at the mouth of the Pachitea and at Masisea, are not believed to be sufficiently consolidated to be classed as anything more definite than partially consolidated clays and sands of late Tertiary or Quaternary age.

A trip made overland from the mouth of the Unini River to the center of the Gran Pajonal (see Fig. 2) failed to throw any additional light on the general geology of the area. Jungle growth and boulder-filled river beds obscure all but a few outcrops. These few, however, substantiate the geology set forth for the Perene and Tambo rivers, and strengthen the theory that the front range scarp is an erosional feature.

COMPARISON OF GEOLOGICAL COLUMNS

In Table I an attempt is made to correlate graphically the formations of the Perene-Tambo rivers with the formations of the Central Andes, Pongo de Manseriche, and Southern Peru. It is impossible to apply the North American time-scale very closely to South American geology. The European scale is more applicable, but even this is unsatisfactory. It has been found advisable on the strength of the fossil determinations of Steinmann,[15] Lisson and Boit,[16] and others,

[14] *Op. cit.*, Vol. XXXIX, pp. 458–60.

[15] *Op. cit.*

[16] C. I. Lisson and B. Boit, *Edad de los Fosiles Peruanos y Distribución de sus Depositos* (Lima, 1924).

to show several of the formations as extending from one period into another, or even from one era to another.

There is a striking similarity in the lithology of the several districts considered, the Cretaceous shale-limestones, cross-bedded sandstones, and the Tertiary red shales being remarkably alike in all of them. Stratigraphically the differences are greater. In the montaña the thick Jurassic and Triassic limestones of the Cordillera are missing, either on account of faulting or because of a major unconformity. Moreover, the Lower Cretaceous sandstone formations are much thicker than in the sierra, and the Basal sandstone of the Cascada formation, which corresponds to the Huacanqui sandstone of the Pongo de Manseriche, is better developed than in the Central Andes. The Cretaceous and the Tertiary formations are quite certainly conformable in the montaña, whereas McLaughlin has found evidence of a marked unconformity at this time in the sierra. Finally, in the montaña the Tertiary tuffs, agglomerates, and intrusives are missing unless the La Merced granite can be considered one of the latter. In their place are found the red shales and the brown sandstones of the Cascada formation.

The stratigraphy reported by Singewald[17] on the Pachitea River is almost identical with that observed on the Perene River, the only difference being that in the latter locality it is impossible to segregate the Brown Beds from the red beds; the latter are apparently overlain directly by the unconsolidated sands, gravels, and clays of Quaternary age, and include the brown sandstones as interstratified members.

Comparison with Bowman's section along the 73d meridian, on the other hand, shows marked dissimilarity. The Tertiary Red Beds, the Lower Cretaceous sandstone-volcanic series, and the Cretaceous lime-shale series which are so conspicuous in all of the stratigraphic sections of northern and Central Peru are not found to the south. Bowman, moreover, remarks on the 1,000–4,000 foot scarp along the eastern edge of the front range, and believes it to be the result of a major fault, the northern extension of a similarly located fault observed by Minchin and himself in Bolivia.[18] As already noted,

[17] *Op. cit.*, pp. 452–60.

[18] "Physiography of the Central Andes," *Amer. Jour. Sci.*, Vol. XXVIII (1909), p. 395.

TABLE I

COMPARATIVE GEOLOGICAL COLUMNS

	Central Andes (McLaughlin)	Perene and Tambo Rivers (Chase)	Pongo de Manseriche (Singewald)	Southern Andes (Bowman)
Quaternary	Glacial deposits and river gravels	Unconsolidated river gravels and clay	Recent alluvium	Recent gravels
Later Tertiary (?)		Partially consolidated red and brown clay and sandstone	Brown Beds	
Early Tertiary (?)	Tertiary intrusives	La Merced granite (?)	Red Beds	Sandstones
	Tertiary volcanics 1800′ est.	Cascada Red Beds; red shale and brown sandstone		
	Tertiary Red Beds 10,000′ est.			Basal red sandstone and conglomerates
		4,000′		
Upper Cretaceous	Jumasha limestone 1500′	Basal sandstone 500′	Huacanqui sandstone	
		Tambo shale	Cretaceous shale	
Lower Cretaceous	Machay limestone 2,000′ est.	Limestone 1,300′–2,200′	Limestone	
	Goyllarisquisga sandstone, and basalt flows 2,600′	Yurinaqui sandstone 10,000′ est.	Pongo sandstone	
Jurassic	Pucara limestone 3,000′			
Triassic	Uliachin limestone			
Carboniferous (?)	Yauli volcanics, Mitu Conglomerates and sandstone 4,500′	Rio Colorado volcanics		Fossiliferous massive, blue limestone with basal conglomerates and sandstone 10,000′ est.
Silurian (?) Ordovician (?)	Excelsior shale			Shales
Pre-Cambrian	Gneisses, schists			Schists

the writer believes that this front range scarp in the vicinity of the Tambo River is due not to faulting but to the effects of erosion upon the soft red shales which overlie resistant sandstones and limestones.

SUMMARY

The particular features which it has been the purpose of this paper to emphasize are:

1. The existence of a major fault zone paralleling the Andean structure, and located just east of the high ranges of the Cordillera Oriental. This fault zone has brought the Carboniferous formations up into contact with the Lower Cretaceous (which represents a vertical displacement of at least 3,000 feet), and probably represents normal faulting of late age along an earlier zone of structural weakness which served as a locus for Tertiary intrusions.

2. The probability that the conspicuous scarp that marks the eastern front range of the Andes is an erosional feature rather than the result of major faulting, and therefore that the younger sedimentary beds extend essentially unbroken, and, save for one localized exception, with only the gentlest folding, all the way from the eastern fault zone at San Ramon to and under the plains of the Amazon basin.

3. The great similarity both lithologically and stratigraphically between the rocks of the high Andes and of the eastern lowlands, indicating that the Andean geosyncline extended at least as far east as the Ucayali River, and north as far as the Pongo de Manseriche. Within this area the formations described in this paper may be expected to occur with but slight variations.

DEPTH CHANGES IN SAGAMI BAY DURING THE GREAT JAPANESE EARTHQUAKE[1]

FRANCIS P. SHEPARD
University of Illinois

ABSTRACT

The sources for the Japanese map showing large modifications of Sagami Bay after the great earthquake have been examined critically. Comparisons of the old and new soundings show that much of the evidence for these changes is of an uncertain character, but that there is good ground for believing that the floors of certain submarine valleys in the bay were decidedly deepened. Other less positive evidence suggests shoaling of one portion of a submarine valley.

The explanation of faulting as a cause of these modifications is unsatisfactory, since the large magnitude of the changes is quite out of keeping with those reported from the bordering lands or from earthquake areas in general. The alternative explanation of submarine landslides appears to account for the phenomena more satisfactorily.

INTRODUCTION

After the great Japanese earthquake of 1923 the Japanese resurveyed the district in which the earthquake was most destructive.[2] The new triangulation showed that a very large area had undergone changes in level. These changes included both elevation and depression, and reached maxima of 8 feet uplift and 5 feet depression. A resurvey of Sagami and Suruga bays was also accomplished at about the same time. On the basis of the old and new soundings the Japanese published maps and sections[3] revealing what they believed to be very considerable modifications of the ocean floor (Fig. 1). In contrast to the changes of a few feet on land, the Japanese reported modifications of the sea floor of as much as 1,300 feet.

SOURCES OF ERROR

Since the largest displacement that has ever been reported as accompanying an earthquake on land was only 47 feet, the changes reported for Sagami Bay have naturally led to much surprise, and some

[1] The writer wishes to acknowledge the co-operation of the Japanese Hydrographic Office in furnishing the information on which most of this study is based. Also, thanks are due to Professor R. R. Kudo for his kindness in translating a Japanese article.

[2] *Bull. Imp. Earthquake Investigation Committee*, Vol. XI, No. 4.

[3] *Earthquake Dist. Surv. Rept., Bur. of Waterways* (September 1, 1924).

doubt has been expressed concerning the validity of the evidence.[4] To be absolutely certain regarding the nature of any changes in the bottom, it would be necessary to know that the soundings were lo-

FIG. 1.—Modifications of Sagami Bay according to the Japanese (Utida). Numbers are in meters.

cated with great accuracy and that the wire was vertical when the soundings were made, and, further, that new soundings were taken

[4] These objections are discussed in part by T. Terada (*Tokyo Imp. Acad. Proc.*, Vol. IV [1928], pp. 45–55); also by C. Davison (*The Japanese Earthquake of 1923* [London, 1931]).

in the same place as the old. Obviously these conditions were not all existent in the Sagami Bay surveys. Regarding location of soundings, it would seem quite possible to make errors of position of from $\frac{1}{4}$ to $\frac{1}{2}$ mile in the center of such a bay. Also, in the deep water there is a chance of the wire being deflected from the vertical by the movements of the ship. This possibility increases with the depth. Apparently no attempt was made by the Japanese to superimpose new soundings on old, but the data of the new survey were sufficiently numerous so that some soundings appear to have hit approximately the same places as the old.

The foregoing considerations show that the soundings cannot give an accurate index of the changes that may have occurred. Whether or not they give grounds for believing that there were changes depends on several factors. The accuracy of the information should vary according to (1) the proximity to land, which increases exactness of location; (2) the spacing of the soundings; (3) the depth of water, which, if shallow, should eliminate errors from inclination of sounding wire; (4) the regularity of the bottom, which, if well developed, should give results despite inaccuracies of locations; and (5) the persistence of changes of the same sign, which tend to show that the differences are not due to local irregularities.

In order to apply these criteria to the supposed changes in Sagami Bay, the soundings of both the old and new surveys were plotted by the writer as accurately as possible on the same tracing cloth. A careful study of the two sets of soundings shows that a large portion of the area from which changes were reported does not have adequate information to fully justify these conclusions. In a large portion of the bay the bottom is so irregular that soundings taken in slightly different localities might be very different. In the northeastern portion of the map the soundings show evidence of increased depth in some places, with evidence of decreased depth in close proximity. Elsewhere in this portion the soundings of the old survey are too scarce to make the comparisons significant. The same general situation exists to the west and northwest of O Shima Island. Here few soundings of the old survey are in close proximity to soundings of the new survey, making comparisons particularly hazardous.

EVIDENCE FAVORING MODIFICATIONS

In spite of this evidence against the conclusiveness of all the reported changes, there appears to be very substantial evidence that changes did occur. Along the course of the deep submarine valley which follows the western side of Sagami Bay the Japanese map shows persistent negative changes. Comparison of the soundings

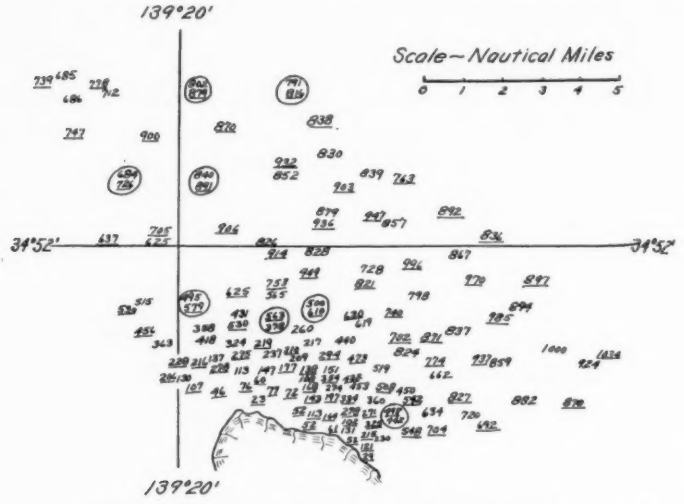

FIG. 2.—Showing the evidence of deepening in a portion of the submarine valley in Sagami Bay. New soundings are underlined. Where two soundings are encircled, the position of both was approximately the same. Soundings in fathoms.

(Fig. 2) shows that this conclusion is amply justified. In an area extending 30 miles north and south and about 8 miles east and west along the axis of the valley, there were thirty-eight pairs of soundings taken in close proximity. All of these showed increased depths in the new survey, with an average of 325 feet and with only seven having less than 150 feet of change. If these apparent changes were due either to errors as regards inclination of wire or to irregularities in the bottom which intervened between the old and the new soundings where they did not hit the same spot, the results would certainly show both negative and positive changes.

Other areas of smaller size contain substantial, but less positive, evidence of modifications. In the submarine valley which emerges from Tokyo Bay, there is an elongate area with four soundings of the old survey ranging from 1,506 to 2,094 feet, which are surrounded and separated by soundings of the new survey ranging from 2,370 to 3,690 feet (Fig. 3). Another depression to the northwest of the last

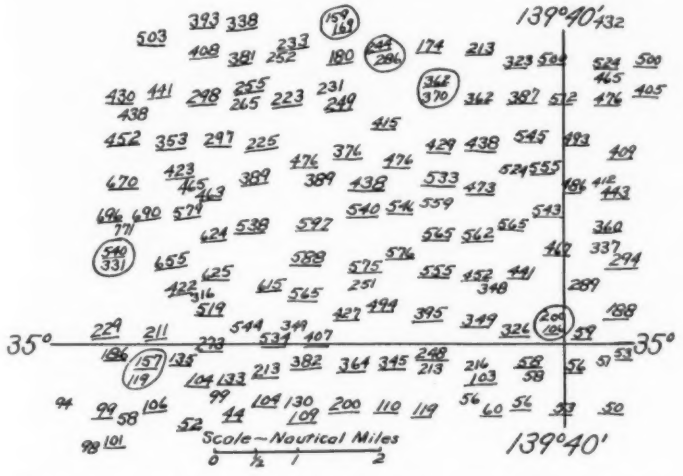

FIG. 3.—Showing soundings of the two surveys just outside the entrance to Tokyo Bay. Note the shallow soundings near the center, surrounded by deeper soundings of the new survey. Symbols the same as in Figure 2. Soundings in fathoms.

shows seven pairs of soundings in close proximity. Of these, six show decrease of depth of an average of 576 feet, and one an increase of 540 feet (Fig. 4).

In addition to the specific evidence of depth modifications mentioned, the soundings in general show much less co-ordination than would be expected if the bottom had not changed or had only changed slightly. That Sagami Bay was deepened generally is suggested by comparison of one hundred and eighteen pairs of soundings of the old and new surveys which were essentially superimposed. While there were only a few more cases suggesting deepening than those suggesting shoaling, sixty-seven to fifty-one, the amount of

deepening averaged 260 feet, as compared to 120 feet for the soundings indicating shoaling. Twenty-seven cases showed deepening of over 300 feet, as against five with shoaling of over 300 feet. In six cases there was shown a deepening of over 600 feet, and in only one an equal amount of shoaling. Comparing soundings of the old and new surveys which were in close proximity, but not superimposed, the evidence in favor of general deepening is still more pronounced.

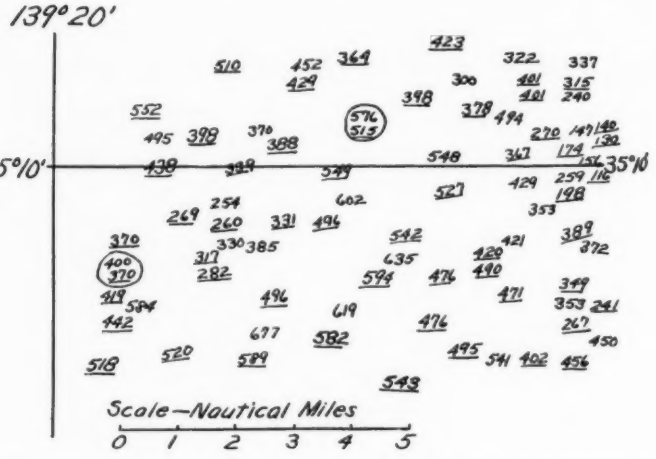

FIG. 4.—Evidence of decreasing depth. Note that the location is at the foot of a slope. Same symbols as in Figure 2. Soundings in fathoms.

TIME OF THE DEPTH CHANGES

Assuming that the evidence given above serves to establish the probability, if not the exact amount, of depth modifications, did the changes all occur at the time of the earthquake? At least 11 years elapsed between the two surveys, so that the changes might have been cumulative. However, if we assume that such shifting was connected with earthquakes, we are confronted with the fact that no other large earthquakes had occurred between the times of the two surveys, in this particular district. Accordingly, it seems reasonable to assume that the changes occurred in a relatively brief time.

FAULTING AS A CAUSE OF THE CHANGES

Most geologists have assumed that the floor of the bay was subjected to fault movements which were large enough to produce these changes. In favor of such a cause Yamasaki showed that north of the deep submarine valley, Sagami River is flowing in a graben of recent origin.[5] These faults may be significant; but if they have been very active recently, it seems strange that there is no estuary at the mouth of Sagami River.

The most serious objection to the fault hypothesis has been mentioned previously, namely, that the displacements have been many times greater than any ever reported on land. If the earthquake had been much larger than any of the land earthquakes, this might be understandable; but seismograph records show clearly that the earthquake was of less intensity than a number of others. Also, the places of greatest shaking were not associated with greatest bottom modifications.[6] Furthermore, the size of the seismic sea waves which accompanied the earthquake were not at all unusual, 43 feet rise being recorded at O Shima as the maximum. A sudden drop of hundreds of feet must surely have produced waves of greater size. Davison suggested that the waves were not greater because the elevation in some places counteracted the depression elsewhere.[7] However, it is difficult to see how this process would have operated at O Shima, since the indications are that a broad depression formed adjacent to it. Also, the area from which evidence indicates elevation is very small compared to that indicating depression. The situation is slightly improved if we consider that the depressions were due to a series of smaller faulting movements.

Another objection to the fault suggestion is that the changes on the lands were not comparable to those on the sea floor. If the changes on the sea floor had all been close to the epicenter, this discrepancy would not have been so significant; but we find that the changes were widespread. Even in Suruga Bay outside of the area of

[5] N. Yamasaki, *Physiographic Studies of the Great Earthquake*, p. 85.

[6] Dr. L. Don Leet informs me that the earthquake was much less pronounced at O Shima Island, where the great deepening occurred, than at the head of the Bay.

[7] C. Davison, *op. cit.*, p. 93.

maximum disturbance[8] there is fairly substantial evidence of changes of the sea floor, especially along the axis of the deep submarine valley which extends out of this bay (Fig. 5).

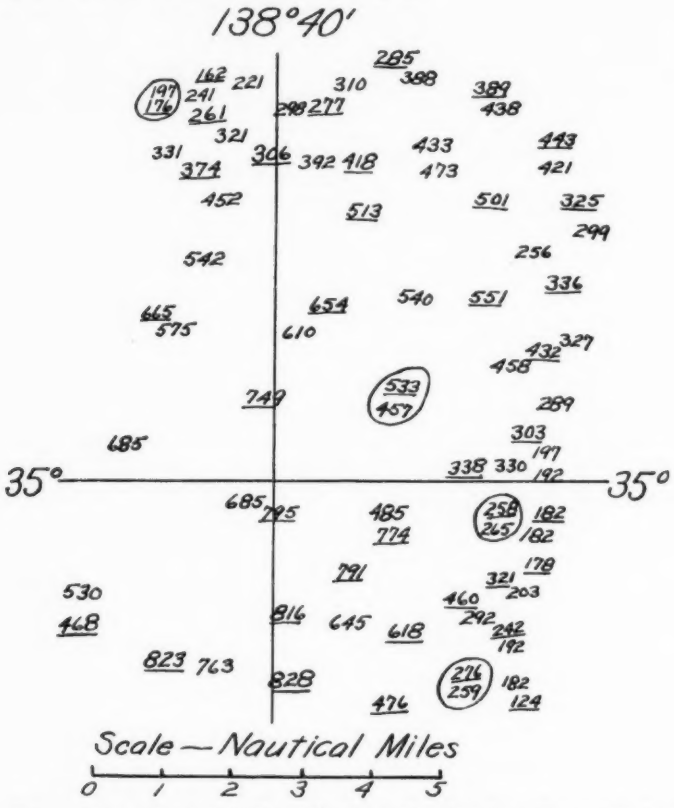

FIG. 5.—Some evidence of depth modification in the submarine valley near the head of Suruga Bay. Same symbols as in Figure 2. Soundings in fathoms.

SUBMARINE LANDSLIDES

Since the evidence does not favor faulting as a cause of the large changes in the sea floor, we may consider the alternative of landsliding. Daly suggested that "some of the changes in depth of the sea bottom may well be due to submarine rock slides or mud slides

[8] For isoseismals see Davison, *ibid.*, Figs. 11, 12, and 17 (from Svda and Imamura).

caused by the earthquake. . . . ."[9] In the same connection Ogawa wrote that the horizontal movement of mud on the sea bottom is easy when tossed by violent shocks.[10] However, Daly adds that most authors have considered that the movements included too much vertical uplift to be due to landsliding. Tujiwhara and Takayama state that the rise of the higher parts of the sea floor and the deepening of the lower make most of the changes impossible of explanation by mud slides.[11] Regarding the rise of the higher portions of the floor of the bay, this argument is not convincing, since examination of the soundings has shown that there is no good evidence for such changes. There remains to explain the deepening of the deeper portions.

The relation of the deepening to submarine valleys in this region appears to be significant. The author has suggested recently that many, or perhaps most, submarine valleys owe their present topography to landsliding.[12] In this case either old fault troughs or possibly drowned river valleys may have received a thick accumulation of fine sediment, perhaps largely volcanic ash. It seems necessary to assume that this fine sediment was at least partly covered by a lava flow, because there are many reports of rock bottom in the bottom of the bay on the old chart. Later, at the time of the great disturbance which centered in Sagami Bay, the saturated sediment underlying the thin crust of lava may have been set in motion and have glided out of the bottom of the bay toward the great Tuscarora deep beyond. It would have accumulated on the floor of the deep, decreasing the depth in that place; but soundings out there are too scarce to make it possible to determine such a change.

The relation to the lands.—Objection to landsliding might be offered on the grounds that there should have been more comparable changes on land. Landslides did occur on land at the time of the earthquake. Davison[13] describes these landslides as having covered a large area to the north of Sagami Bay. The slides did not produce changes comparable with those in Sagami Bay, since they were prin-

[9] R. A. Daly, *Our Mobile Earth*, p. 71.

[10] T. Ogawa, *Chikyu* ("The Globe"), Vol. I (1924) p. 405.

[11] S. Tujiwhara and T. Takayama, "Mechanics of the Great Sagami Bay Earthquake," *Bull. Earthquake Res. Inst. Tokyo Univ.*, Vol. VI (1929), p. 178.

[12] "Submarine Valleys," *Geog. Review*, Vol. XXIII (1933), pp. 77–89.

[13] *Op. cit.*, pp. 106–8.

cipally connected with dislodgment of material from steep slopes and did not include mud flows of sediment from valley fills. There are several reasons why slides on land would have been different from those on the sea floor. The valley fills on land were probably much smaller than those within the bay, the sediment on the sea floor was more completely saturated, and the sediment was very likely finer, containing more colloidal material and thus more subject to sliding.

Detailed nature of the bottom.—The soundings of both the old and the new survey were too limited to indicate the exact nature of the submarine valley in Sagami Bay. It would be interesting to make echo-sounding profiles across this area in order to obtain more complete information. At present the soundings are sufficient to show that there are some hills in the floor of the valley such as would be expected to be left from a swirling mud flow.

DECREASE OF DEPTH IN DEPRESSIONS

There remains to be explained the reason for the decrease in depth which is suggested in Figure 4. Examination of the ocean floor around this area shows that the decrease in depth is practically confined to a submarine valley. Such changes may be explained by the dislodgment of sediment from the valley walls at the time of the earthquake. The result of this process would be to deepen the water on the sides of the valley and shoal the water in the valley bottom. It is much easier to determine changes of depth on a valley floor than on the steep slopes of the sides, so that the shoaling in this case would be more evident than the accompanying deepening where the material was dislodged.

LANDSLIDES AND SEISMIC SEA WAVES

The largest recorded sea waves accompanying the earthquake were on the island of O Shima, where they rose to heights of 13 meters. The waves were not nearly so large along the coast at the head of the bay. As stated previously the epicenter of the earthquake was near the head of the bay, whereas the changes in the sea bottom appear to have been most pronounced east of O Shima. If these changes in the bottom were due to landsliding, were not the sea waves caused or augmented by the same phenomenon? The movement of a large mass of material along the bottom must drag the water along with it and set up violent surface disturbances.

ALTERATION OF A COLORADO GRANITE TO SERICITE SCHIST

G. M. SCHWARTZ
University of Minnesota

ABSTRACT

Petrographic and chemical study of fresh granite, sericitized granite, and sericite schist shows the changes characteristic of this type of alteration. The principal chemical changes were a large loss of soda and a gain of water. The difference between the sericitized granite and the sericite schist is principally in texture and structure.

INTRODUCTION

During a reconnaissance in 1929 of some abandoned mines west of the Leadville district the writer was shown several phases of altered granite by Mr. E. P. Chapman, geologist of Leadville, Colorado. These rocks seemed to show exceptionally well many phases of alteration from granite to sericite schist. Examination of thin sections of two of these rocks later confirmed their character, and in 1930 the writer, through the courtesy of the E. J. Longyear Company of Minneapolis, was given time to examine the occurrence more in detail and collect a suite of specimens for microscopic investigation and chemical analysis.

Dr. C. H. Behre, Jr., and Mr. A. E. Sandberg visited the occurrence and discussed the relations with the writer. Dr. Frank F. Grout kindly reviewed the microscopic work.

Examples of sericitized granites are common in connection with mineralization, but places where sericitization has been followed by shearing, thus producing a schist, are not often found and therefore this occurrence is described in detail.

LOCATION

The dump and the tunnel in which the altered granite is exposed is on the north side of Lake Creek, a tributary of the Arkansas River directly across the valley from Leadville, Colorado (see Fig. 1). The tunnel is on the Gold Plate No. 1 claim a short distance above the more extensive workings known as the Hap Hazard Mine. This is

almost 2 miles above Turquoise Lake, which is formed by a dam across Lake Creek. The area lies on the northeast flank of Mount Massive, the fourth highest peak in the United States. Mount Massive is a part of the Sawatch range which forms the continental divide from Tennessee Pass southward in Colorado.

GEOLOGIC SETTING

Although the area referred to in this paper lies just across the valley from the Leadville district, it is not considered a part of that

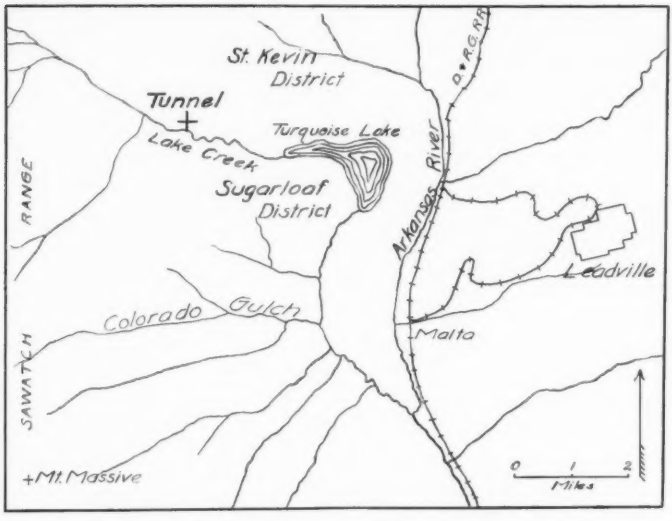

FIG. 1.—Sketch map showing the location of the tunnel from which the sericitized granite was obtained.

district and is not described in the monograph and professional paper of the United States Geological Survey on the Leadville district. The area is, however, shown on the Leadville topographic sheet.

The rocks of the area discussed consist mainly of pre-Cambrian granite with smaller amounts of Tertiary porphyries. Dr. C. H. Behre, Jr., correlated the granite with the Silver Plume granite of other pre-Cambrian areas in Colorado. This granite is very widespread and in general seems to form the main part of the Sawatch

range from the Mount of the Holy Cross south to Twin Lakes where large masses of Tertiary quartz monzonite intrude it.[1]

In the Sugarloaf–St. Kevin area (Fig. 1) this granite is cut by extensive shear zones and over wide areas is so completely sericitized that it superficially resembles some of the Tertiary porphyries of the Leadville district. Silicification is important near the veins and pyrite is a common associate of sericite. The veins range from mere stringers to a width of several feet, and some extend for a length of a thousand feet or more.

Two types of ore deposits were formed in connection with the mineralization. The first is a complex lead-zinc-silver-gold ore oxidized near the surface, but passing into sulphide at depth. The second type of ore is a siliceous pyritic silver-gold ore in which, according to those familiar with the district, the values were best where small amounts of tetrahedrite occurred with the pyrite.

Microscopic studies suggest that the complex sulphide veins represent two phases of mineralization, as a late generation of quartz-pyrite veinlets cuts the earlier pyrite-sphalerite-galena ore. A detailed study of the paragenesis of the sulphide ore of the Sugarloaf area will be published by Mr. A. E. Sandberg.

The mineralization has many features in common with that of the Leadville district, and is presumed to be of the same age, i.e., early Tertiary. The solutions were doubtless furnished by the great series of Tertiary intrusives of the region.[2]

In the vicinity of the tunnel noted above the most detailed information is available from the Hap Hazard workings, which, according to Chapman,[3] lie in a brecciated and much silicified area of the granite. Silification and pyritization are important, and the rock resembles that of the molybdenum deposits at Climax about 10 miles to the northeast. Cutting through the brecciated area are several narrow but persistent veins which in places carry good values. In some veins the value is principally due to gold. In others silver is predominant. Sericitization is evidently related to the many shear zones

[1] R. D. Crawford, "A Contribution to the Igneous Geology of Central Colorado," *Amer. Jour. Sci.*, Vol. VII (1924), pp. 365–88.

[2] *Ibid.*

[3] E. P. Chapman, personal communication.

of the region. There is some evidence of a chimney-like area of mineralization, but more detailed work is needed on the problem.

OCCURRENCE OF THE SERICITIZED AND SHEARED GRANITE

As described above, highly sericitized granite occurs over a wide area. In the particular occurrence described here additional interest was furnished by the fact that locally the sericitized granite seemed to have been sheared to such an extent that it is now a sericite schist. Considered by itself, one would be justified in wondering whether it ever was granite. A short distance upstream from this occurrence the granite occurs as large outcrops apparently little affected by hydrothermal solutions and only slightly by weathering. Thus three phases of the granite are distinguished: (*a*) relatively unaltered granite, (*b*) highly sericitized granite, and (*c*) sericite schist. Specimens of relatively fresh granite were also obtained in Colorado gulch along the south side of the Sugarloaf district.

The tunnel in which the specimens of sericitized granite and sericite schist were obtained is about 140 feet long. There are some inclusions of basic schist in the granite and several shear zones with a small amount of gouge. The sericite schist seems to grade into rock which retains the granitic texture; in fact, this gradation may be observed in the hand specimens. There are also fine-grained phases of the sericitic rock which were puzzling in the field and under the microscope as well. The best opinion seems to be that these were possibly quartzitic inclusions in the granite. These rocks were observed in place and also on the dump.

MICROSCOPIC DESCRIPTION

Unaltered granite.—The relatively unaltered granite selected for comparison with the sericitized phases outcrops a few hundred feet up the valley from the tunnel in which the sheared material occurs. The rock is a biotite granite with a medium-grained texture (Fig. 2). Feldspars are microcline, orthoclase, and albite and are somewhat altered to sericite and dusty kaolinite. Biotite is abundant and is partly altered to chlorite resulting in a green color. Muscovite is abundant and small amounts of leucoxene and limonite were also noted. Quartz makes up about 20 per cent of the rock.

The fresh granite of Colorado gulch is essentially of the same type. It consists of 20 per cent quartz. Feldspars are microcline, orthoclase, and albite, with biotite and muscovite as essential minerals and magnetite, apatite, and titanite as accessory minerals. The feldspar is somewhat clouded with sericite, and biotite is partly altered to chlorite.

Sericitized granite.—Thin sections of several sericitized granite specimens from the tunnel and from the Sugarloaf area were examined. The one selected for analysis was collected at the portal of the tunnel and shows the granitoid texture in the hand specimen so perfectly that there is no doubt as to its original nature. In hand specimen this rock is light gray in color, relatively soft, with the gray quartz grains clearly distinguishable from the dead white of the sericite.

Quartz and sericite are the principal constituents. The quartz is conspicuously shattered, but otherwise is probably not much changed from its original character. It composes about 25 per cent of the rock. The feldspar areas have been almost completely sericitized (Fig. 3) with the freeing of some quartz. At places shadows of the feldspar twinning remain in the sericite. Clear muscovite areas are presumably original in the granite; other muscovite plates are dusty with rutile or leucoxene and are interpreted as secondary after biotite.

Sericite schist.—The sericite schist is a gray rock having well-developed schistosity along parallel zones with more granitic material between (Fig. 4). Pyrite is abundant in some specimens, occurring as films on the rock-cleavage surfaces. The principal minerals are quartz and sericite. The quartz occurs as irregular grains, but little changed from the original granite. The feldspar areas of the granite have been converted to a mat of sericite which is coarse grained along the schistose bands. In the most schistose zone, quartz is shattered and the fractures filled with sericite. Pyrite is abundant and occurs mainly in muscovite areas which were presumably derived from biotite. Leucoxene is fairly abundant and contains rutile needles which have been formed by recrystallization.

Fine-grained sericite rock.—Occurring with the rock types described above are certain fine-grained sericite rocks whose origin was

FIG. 2

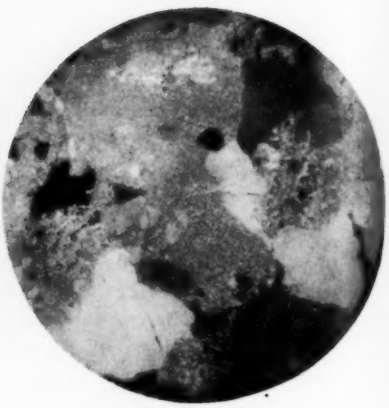

FIG. 3

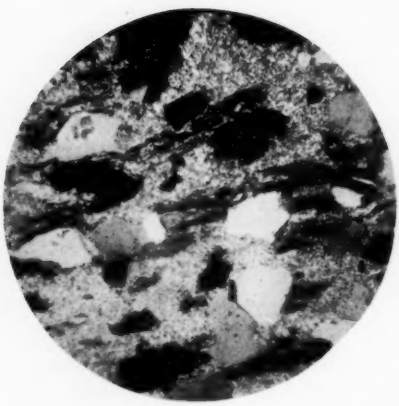

FIG. 4

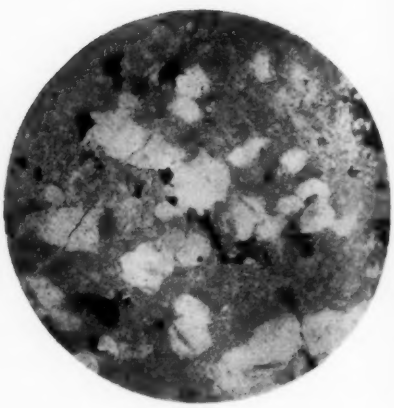

FIG. 5

FIG. 2.—Photomicrograph of fresh granite from Lake Creek Valley. Same rock as analysis 1. Crossed nicols ×15.

FIG. 3.—Photomicrograph of sericitized granite from portal of tunnel. Same rock as analysis 2. Quartz areas remain essentially unchanged. Feldspars are almost completely sericitized. Plain light ×15.

FIG. 4.—Photomicrograph of sericite schist derived from granite. End of tunnel. Same rock as analysis 3. Crossed nicols ×60.

FIG. 5.—Photomicrograph of fine-grained rock probably sericitized quartzite inclusions in granite. Dump of tunnel. Plain light ×20.

not considered certain from the field relations. They might have been either finer-grained granite or inclusions so completely changed as to mask their original character. Thin sections of these rocks show the principal constituents to be quartz, sericite, and muscovite with lesser leucoxene, rutile, and pyrite. The texture of the altered rock suggests that the rock is an altered quartzite. The quartz gives a distinct impression of having been sedimentary grains (Fig. 5). It seems probable that this rock was originally an arkosic quartzite, but the possibility of its being an aplite is not excluded.

CHEMICAL CHANGES

Complete rock analyses[4] of the fresh granite, sericitized granite, and sericite schist were made by Dr. R. B. Ellestad of the Rockefeller National Laboratory for Rock Analysis at the University of Minnesota (see Table I).

TABLE I

CHEMICAL ANALYSIS

	1*	2†	3‡		1*	2†	3‡
SiO_2	72.80	78.09	62.39	H_2O+	.86	1.82	2.52
Al_2O_3	14.82	13.71	18.94	TiO_2	.20	.18	.91
Fe_2O_3	.37	.56	.82	P_2O_5	.23	.09	.19
FeO	.89	.24	.34	S	Tr.	.04	.....
MgO	.45	.52	.72	MnO	.02	Trace	Trace
CaO	.59	.11	.11	FeS_2	.....	.....	6.70
Na_2O	3.00	.12	.46				
K_2O	5.63	4.05	5.70	Total	99.98	99.63	99.94
H_2O-	.12	.10	.14				

* Fresh granite. † Sericitized granite. ‡ Sericite schist.

Examination of the analyses shows that the sericitized granite and sericite schist are much alike chemically except for the large content of iron sulphide and higher titanium in the sheared rock. The pyrite is not limited to the sheared rock in general, as was shown by study in the field and examinations of specimens other than those analyzed. Pyrite tends to be more abundant in the sheared rock probably because of greater freedom of circulation. There is, there-

[4] The analytical work was made possible by a grant from the research funds of the graduate school of the University of Minnesota.

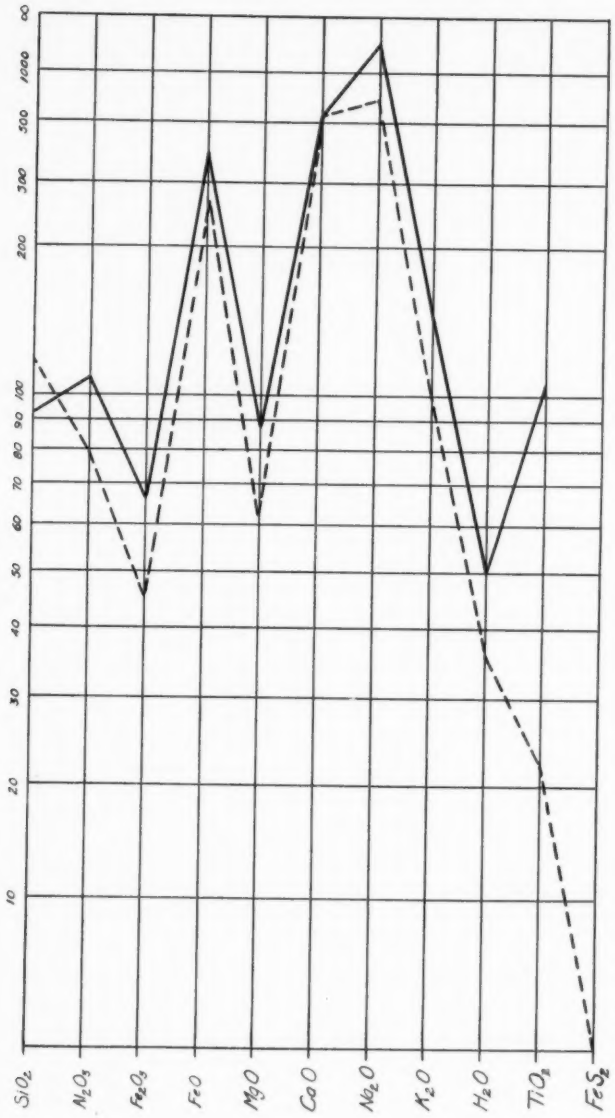

FIG. 6.—Straight-line diagram representing gains and losses of constituents in altered as compared with fresh rock. Solid line, sericitized granite; dash line, sericite schist.

fore, no important consistent difference in composition between the two types of altered rocks.

The altered rocks as compared with the unaltered granite show a striking loss of soda and a gain in water. The potash was retained while soda was almost entirely eliminated.

In Figure 6 a straight-line diagram[5] is used to show graphically the changes in chemical composition of the altered rocks as compared with the fresh granite. The close correspondence of the change in the two altered phases is clearly shown.

Comparison with the composite diagram for gains and losses of constituents in the katamorphism of acid igneous rocks by thermal solutions given by Leith and Mead[6] indicates that the changes in the Lake Creek granite are typical of hydrothermal alteration of acid igneous rocks.

SUMMARY

1. A pre-Cambrian granite was altered to a sericitized rock and field relations show that at places this rock was sheared to form a sericite schist.

2. Microscopically the sericitized granite shows the feldspars almost completely altered to sericite with the freeing of some quartz, or possibly quartz was added during alteration. The original quartz grains remain essentially unchanged.

3. The sericite schist shows much larger sericite blades than the sericitized granite and this coarse sericite grades into platy muscovite. The quartz grains are fractured and the fractures filled by sericite needles.

4. Chemical changes during alteration were mainly a large loss of soda and a gain of water. Pyrite was added in important amounts to the schist and in some cases to the sericitized granite. Titanium seems to have been added to the sheared rock. The potash of the original rock was retained, but the analyses indicate no addition during alteration.

5. The essential difference between the sericitized granite and the sericite schist is in texture and structure.

[5] See C. K. Leith and W. J. Mead, *Metamorphic Geology* (New York: Holt & Co., 1915), p. 12, for method of plotting.

[6] *Ibid.*, Pl. V.

THE JASPER CONGLOMERATE, AN INDEX OF DRIFT DISPERSION

CHESTER B. SLAWSON
University of Michigan

ABSTRACT

The source of Jasper conglomerate erratics of the central Great Lakes region is shown to be a small area of outcrops lying about 30 to 40 miles east southeast of Sault Ste Marie. The limits of distribution as determined by the available data are outlined for both the Wisconsin and the Illinoian drift. The greater part of these erratics were transported through the Saginaw ice lobe, although the eastern half of the Lake Michigan lobe deposited many of them, especially during the Illinoian ice movement.

In the material that comprises the glacial drift of the central Great Lakes region there is no boulder type so characteristic and so easily identified as the Red Jasper conglomerate. Because of its striking association of bright red and dark Jasper with translucent quartz pebbles in a background of pure white quartzite, these erratics have been used in many sections as ornamental stones. The source of the conglomerates has long been known to be the original Huronian sediments described by Logan and Murray. These sediments extend from the shore of Lake Superior east along the north shore of North Channel and Georgian Bay where they swing north through the Sudbury mining district. Logan and Murray described three areas in which Jasper conglomerates occur—along the Sturgeon, Wanapitei, and Whitefish rivers (north shore of Georgian Bay),[1] the Bruce Mines district,[2] and Goulais Bay (Lake Superior).[3] Recent investigations[4] have correlated this Huronian area with the better-known areas to the east and north in which the succession has been worked out by Coleman and Miller. Logan and Murray's

[1] *Geology of Canada* (1863), pp. 51–52; see also map facing p. 166; *Geol. Surv., Can. Rept. of Prog.* (1855) (Wahnapitae = Wanapitei).

[2] *Geology of Canada* (1863), p. 57.

[3] *Ibid.*, p. 63.

[4] T. T. Quirke, and W. H. Collins, "The Disappearance of the Huronian," *Geol. Surv. Can. Mem. 160* (1930), p. 28; W. H. Collins, "North Shore of Lake Huron," *ibid., Mem. 143* (1925), pp. 67–69; R. G. McConnell, *Ont. Bur. Mines Ann. Rept.*, Vol. XXXV (1926), Part II, pp. 27–30.

Red quartzite, Red Jasper conglomerate, and White quartzite formations, whose type area was north of Bruce Mines, are now grouped as the Lorrain quartzite of the Cobalt series which extends in a series of intermittent outcrops from Lake Superior east through the entire North Shore district.

The Lorrain carries through the whole region a horizon of quartz conglomerate of variable thickness in which Jasper pebbles are more or less prevalent. This horizon serves as a means of correlating the various outcrops. Because of the diagnostic character of the Jasper, the terms "Jasper conglomerate" and "Red Jasper conglomerate" have been applied to any rock carrying an appreciable amount of Jasper pebbles. Thus some confusion has arisen as to the source of the erratics found in the drift. Among glacialists[5] the source of these boulders is commonly believed to be the north shore of Georgian Bay. To account for their distribution along the western edge of the Illinois drift in Lee County, Iowa, Leverett[6] has postulated two steps in their transportation: an earlier ice movement with a strong westerly component strewing boulders along its path, followed by the Illinoian ice movement which carried them to their present location. It is the purpose of this paper to show that what is commonly referred to as the ornamental type of Jasper conglomerate has its counterpart 100 miles west of Georgian Bay in the vicinity of the Thessalon River, north and northeast of Bruce Mines, Ontario, approximately 30–40 miles east southeast of Sault Ste Marie (Fig. 1). This brings the source of these boulders close to the axial position of both the Lake Huron and the Lake Michigan lobes.

The misconception in regard to the origin of these boulders probably arose from the descriptions of Logan. In his discussion of the Pleistocene of Canada he was the first to use these erratics to plot the direction of ice movement. He says,[7] "Bowlders of jasper conglomerate occur on Lacloche Island, and on Fox Islands, thirteen miles west of French River, but have not yet been observed on the coast east of the river." Lacloche Island separates Georgian Bay from North Channel while the Fox Islands lie off the north shore of

[5] F. Leverett, "The Illinois Glacial Lobe," *U.S. Geol. Surv. Mon. XXXVIII*, pp. 24, 41, 394; also *Mon. XLI*, pp. 421–22, and *Mon. LIII*, p. 117.

[6] *Ibid.*, *Mon. XXXVIII*, p. 394.

[7] *Op. cit.*, p. 894.

the bay proper. Murray[8] who described the formations lying immediately north of these islands from which these boulders were

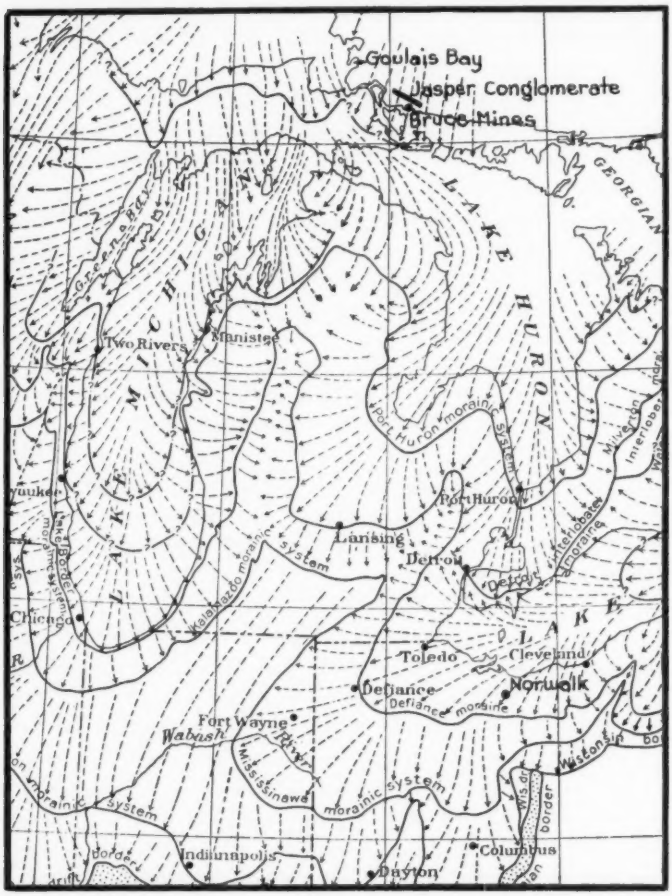

FIG. 1.—Base map taken from Leverett, *U.S. Geol. Surv. Mono. LIII*, Plate V, facing page 62.

derived, describes the rocks as white to greenish-white quartzites and conglomerates, and says, "The pebbles of the conglomerate are

[8] *Geol. Surv., Can., Rept. of Progress for the Year 1855*, pp. 173–76; see also *Geology of Canada*, pp. 51–52.

chiefly small white opaque rounded masses of quartz, but these are occasionally mixed with masses of red and green jasper." Collins[9] describes these same formations and states, "There are almost no jasper pebbles in it, ..." and again,[10] ". . . . is an uninteresting rock of light grey or pink color, consisting of white and gray quartz pebbles in a quartzite matrix; but towards Sault Ste Marie pebbles of dark grey and bright red jasper become more abundant." Although this area has not been visited by the writer the evidence of these reports seems concrete enough to rule out the Georgian Bay area as a source of these boulders.

In examining the western part of this Huronian area, consideration was given not only to the outcrops, but also to the character of the drift in the vicinity of the outcrops. North of Sault Ste Marie the Lorrain is represented by erosional remnants[11] in ridges on the north shore of Goulais Bay, in King Mountain 20 miles due north of Sault Ste Marie, and in Bellevue ridge 16 miles north-northeast of Sault Ste Marie. These conglomerates as exposed are not of the highly ornamental phase to which the term "Jasper conglomerate" has been applied in the drift of Michigan, but contain pebbles of yellow brown and red Jasper with some interstitial mica. While distinct from the conglomerates of the Georgian Bay district, they are probably comparable to it in general abundance and size of the Jasper pebbles.

It may be argued that this area was a source of typical Jasper conglomerate boulders, for, in the upper portions of the Lorrain formation which have been eroded, there may have been a conglomerate of the ornamental phase. If this were true the logical assumption is that the drift immediately south of these outcrops should contain boulders of that character. Salisbury[12] estimated that at least 70 per cent of the drift material in an area of outcrops is of local derivation. The character of the present drift then, in many respects, is a better criterion of the type of glacial material derived from an area than the outcrops now existing in that area. In studying the composition of the drift east of Lake Superior, boulders of the type under

[9] *Op. cit.*, p. 69.

[10] *Ibid.*, p. 135.

[11] R. G. McConnell, *Ont. Bur. Mines, Ann. Rept.*, Vol. XXXV (1926), Part II, p. 28.

[12] R. D. Salisbury, "The Local Origin of the Drift," *Jour. Geol.*, Vol. VIII (1900), p. 429.

consideration were not found until a point a few miles east of Echo Bay was reached. East of Desbarats they are very abundant. The village of Bruce Mines has erected in the school yard a memorial to war veterans composed of typical Jasper conglomerate blocks all cut from erratics. Good exposures of the outcrop may be seen in the N.E. $\frac{1}{4}$ of Lot 11, Concession I, Aberdeen Twp.; Lot 3, Concession IV, Plummer Twp.; Sec. 28, Rose Twp.; and Sec. 12, Lefroy Twp. When plotted on the map these outcrops lie in a straight line, 13 miles in length, striking N. 125° E., roughly paralleling the Murray fault 3 miles to the south. The formation dips rather uniformly toward the fault with an average dip of 25°. The topography is rather rough, consisting of ridges and knobs developed by pre-Pleistocene erosion, with the valleys partially filled with Pleistocene deposits.

The lowermost part of the formation (Murray's Red quartzite) is a red feldspathic quartzite which grades upward into a pure white quartzite matrix carrying translucent quartz and Jasper pebbles (Murray's Red Jasper conglomerate). The conglomerate phase develops gradually somewhat before the matrix loses its reddish character. The Thessalon River flows west out of Rock Lake between high ridges of conglomerate characteristic of this transitional phase. As the conglomerate phase develops, the matrix becomes pure white, the pebbles increase in size up to $2\frac{1}{2}$ inches, and the rock takes on the highly ornamental character described by Murray[13] as, "The pebbles . . . . are sometimes arranged in thick bands, and blood-red Jaspers often predominating in a nearly pure white base produce a brilliant, unique, and beautiful rock." The conglomerate represents approximately the middle third of the formation, the upper third (Murray's White quartzite) being a fine, dense white quartzite.

The southwest side of the Murray fault is upthrown with a vertical displacement of over 6,000 feet. This duplicates the Huronian formation to the southwest, but this area is relatively heavily covered with drift. One outcrop corresponding to the basal portion of the Jasper conglomerate horizon was seen in the low ridge running north and south in Lot 5, Concession I, Tarbutt Twp. Other report-

[13] *Geology of Canada*, p. 57.

ed outcrops in this area lie on the north shore of St. Joseph Channel, on the east shore of Lake George, and on the north shore of Desbarats Lake.

The phase of the Jasper conglomerate, which has been so termed in the drift erratics, is thus delimited to an area 20 miles wide in an east-west direction. Its north-south extent is left indeterminate, but as this is in the general direction of the ice movement it will affect but little any conclusions drawn relative to drift dispersion. Murray[14] reported an outcrop of the basal phase of the Jasper conglomerate on the east shore of Lake Wakomata, 30 miles northeast of Bruce Mines, and there is an unverified report of Coleman[15] of an outcrop north of Echo Lake. The probability that a rock so unusual in character could maintain that charater over any considerable distance is very small.

In the Wisconsin drift the eastern limit of distribution of these erratics is marked in Michigan by the moraines bordering the glacial lake beds and in Ohio by a line running south of Norwalk through Mansfield and east of Dayton. On the west a limited number of the boulders have been deposited by the eastern half of the Lake Michigan lobe. In general, though, the great bulk of them were transported by the ice of the Saginaw lobe. In crossing the interlobate tract between the Saginaw and the Lake Michigan lobes one observes a marked diminution in the prevalence of these boulders when the drift of the Lake Michigan lobe is entered. The eastern limit of distribution in the Illinoian drift corresponds closely to that of the Wisconsin drift as evidenced by erratics in the vicinity of Cincinnati.[16] To the west the Illinoian drift carried the Jasper conglomerate as far as the extreme southeastern corner of Iowa.[17] What little evidence is available from field observations southwest of Lake Michigan indicates that the Illinoian ice movement through the Lake Michigan basin carried a higher percentage of these erratics than the Wisconsin movement.

[14] *Ibid.*, p. 842. (Wahcomatagaming = Wakomata.)

[15] *Ontario Bur. Mines, Ann. Rept.*, Vol. VIII (1899), pp. 124 and 159.

[16] F. Leverett, *U.S. Geol. Surv., Mon XLI*, p. 261.

[17] *Ibid., Mon. XXXVIII*, pp. 24, 41.

In regard to the more local distribution, especially within the Saginaw lobe, the occurrence might be termed highly irregular. One morainic ridge may contain a relatively high percentage of Huronian rocks while an adjacent ridge may be nearly devoid of them. There is a definite association of Jasper conglomerate with other Huronian rocks, chiefly quartzites, which are typical of the Bruce Mines area. While the conglomerate is conspicuous because of its striking appearance, it will in no case represent as much as 1 per cent of the boulders of an area.

CHEMICAL CHANGES IN METABASALT FROM SOUTHERN QUEBEC[1]

H. W. FAIRBAIRN
Harvard University

ABSTRACT

This paper presents two new chemical analyses of metabasalt from Southern Quebec and shows that the extensive mineralogic changes involved in uralitization and later ankeritization of basalt are accompanied mainly by addition of H_2O and CO_2. This is characteristic of uralitization, but to the writer's knowledge has not before been described in the case of ankeritization. There is very little loss or gain of bases during the alteration, and the process is endomorphic, probably taking place in a temperature range between that required for uralitization and calcite-chloritization.

INTRODUCTION

The chief purpose of this paper is to place on record two chemical analyses of metabasalt in Southern Quebec which indicate the trend of its metamorphism. This is supplemented by a brief outline of the occurrence, field relations, petrography, and a comparison of the alteration with other localities showing similar features.

OCCURRENCE

The metabasalt studied by the writer is situated along the west side of Lake Memphremagog about 75 miles southeast of Montreal and occupies an area of approximately 100 square miles. It forms part of a larger body of flows which extend northeast parallel to the regional strike of the schists and slates.[2] The outcrops are exposed mostly as isolated hills and ridges having steep slopes and a maximum relief of 2,000 feet.

FIELD RELATIONS

The metabasalt forms part of the Bolton igneous group recently described by the writer and T. H. Clark.[3] It rests unconformably on

[1] Published with the permission of the Director, Geological Survey, Department of Mines, Ottawa, Canada. Study made in the Department of Mineralogy and Petrography, Harvard University.

[2] J. A. Dresser, "Igneous Rocks of the Eastern Townships of Quebec," *Bull. Geol. Soc. Amer.*, Vol. XVII (1906), p. 497.

[3] Can. Geol. Surv. (unpublished memoir).

Upper (?) Cambrian and Middle Ordovician slate and quartzite and is overlain unconformably by Middle Silurian conglomerate, shale, and limestone. All the sedimentary rocks of the district are folded, and those of pre-Silurian age especially are schistose or slaty, and recrystallized. The metabasalt is considered to be late Ordovician and to represent part of the igneous activity which accompanied the Taconian orogeny of northeastern United States and Canada. The dip of the flows is known only in one or two places and is always steep. Nothing is known of the succession or thickness of individual flows, but there is reason to believe that a total of several thousand feet is present.

The metabasalt forms two distinct lithologic types known as uralite rock and ankerite rock. The latter is irregular in distribution and the largest outcrop is about 300 feet across. The uralite rock composes much the greater part of the metabasalt and grades into the ankerite type. Both forms show evidence of their original extrusive character in abundant pillow structure, fine texture, and common amygdules.

Metagabbro dikes, usually less than 30 feet, are associated with the metabasalt, and in a few places intrude the Ordovician slate. They are considered to be feeders to the flows, but are of minor importance areally.

PETROGRAPHY

Uralite rock.—The uralite rock is typically grey-green, massive, and fine textured. Under the microscope the essential minerals are seen to be albite-oligoclase, actinolitic amphibole, epidote-clinozoisite, chlorite, and leucoxenized titanite. The albite-oligoclase is in small, irregular, clouded grains which in places show broad albite twins indicative of a more calcic primary plagioclase. Actinolite occurs in abundant minute shreds and flakes and gives the effect of a fine-matted texture. Blue clinozoisites are more common than ordinary epidote and occur as minute grains full of inclusions. Chlorite commonly replaces actinolite and is found in disseminated flakes or small segregations. Titanite is abundant in minute, irregular grains and is usually clouded by whitish leucoxene. Accessory minerals include calcite in small, irregularly distributed grains, quartz in nested aggregates of minute, polygonal crystals, and sericite in very small

shreds on the plagioclase. Pyrite and pyrrhotite are of regular occurrence. Magnetite and apatite are uncommon. Primary minerals, olivine and green hornblende, occur as phenocrysts in a few localities. They are embedded in a typical uralite groundmass and are themselves partly altered.

Ankerite rock.—The ankerite rock is usually buff colored, massive, and of fine texture. It is most easily recognized in the field by its rusty, weathered surface, due to decomposition of the carbonate and formation of hydrous ferric oxides. Under the microscope the essential minerals are seen to be albite-oligoclase, ankerite, quartz, and chlorite. The plagioclase is similar to that described for the uralite rock. Ankerite rhombs with perfect crystal boundaries are the characteristic feature of these rocks and usually form the largest grains in the sections. Quartz is more common than in the uralite rock and occurs in irregular segregations. Chlorite is found in disseminated patches of small flakes. Sericite is the most prominent accessory and occurs as minute shreds on the plagioclase. Apatite, rutile, and pyrite are found in small amounts. Epidote is uncommon, and no actinolite is found in any of the sections examined.

CHEMICAL ANALYSES

Chemical analyses of the uralite and ankerite phases of the metabasalt are shown in Table II. The samples for analysis were collected from many representative specimens of each of these rocks, and the analyses therefore give approximate average compositions. Comparison of the uralite rock analysis with Daly's average basalt shows an undoubted relationship which justifies naming the rock a metabasalt. This is in complete accord with comparisons made of chemical analyses of many altered basalts throughout the world and furnishes one more example of the small change in chemical composition which accompanies widespread change in mineral composition.

The relation of the ankerite to the uralite rock analysis is even more significant. Recalculation on the basis of CO_2 equivalent to that found in the uralite rock reveals a striking parallelism in the percentages of all constituents. No other recalculation would preserve this exact balance and indicate the same relation to the uralite rock analysis. This feature has, therefore, particular importance and indi-

cates that there was no essential loss or gain of bases during carbonatization of the uralite rock and that addition of CO_2 was the only notable change during this stage of metamorphism.

TABLE I

TABLE OF ANALYSES*

	Average of 198 Basalts (Daly)	Uralite Rock	Recalculation of Ankerite Rock	Ankerite Rock
SiO_2	49.06	49.44	48.23	42.64
TiO_2	1.36	1.43	1.17	1.04
Al_2O_3	15.70	15.92	16.24	14.42
Fe_2O_3	5.38	2.33	2.53	2.25
FeO	6.37	8.11	8.35	7.41
MnO	0.31	0.10	0.08	0.07
MgO	6.17	6.10	6.79	6.03
CaO	8.95	8.20	8.10	7.20
Na_2O	3.11	3.06	2.98	2.65
K_2O	1.52	0.50	0.46	0.41
P_2O_5	0.45	n.d.	.....	n.d.
CO_2	.....	1.31	1.31	12.52
H_2O	1.62	3.40	2.91	2.58
	100.00	99.90	99.15	99.22†

* Analyses by F. A. Gonyer, Department of Mineralogy, Harvard University.

† Organic matter present.

APPROXIMATE PERCENTAGE OF MINERALS*

Mineral	Uralite Rock	Ankerite Rock
Albite	23	20
Chlorite	20	20
Actinolite	20	..
Epidote } Clinozoisite	20	0.5 ..
Quartz	5	20
Calcite	4	..
Ankerite	..	34
Sericite	4	3
Titanite	3	1
Others	1	2

* Albite, calcite, ankerite, titanite, and sericite calculated from chemical analyses. Others estimated from thin section. In the text albite is referred to as albite-oligoclase as some optical measurements indicate a small percentage of the lime molecule. The optical data in general, however, indicate pure albite and for simplicity in the feldspar calculation only the soda molecule is considered. The lime molecule is therefore included with the epidote, ankerite, etc., but is not believed to be large enough to affect the proportions of these minerals.

EXAMPLES OF ANKERITIZATION

Chemical data concerning carbonatization are largely restricted to areas of economic interest. In non-economic areas, such as the one described in this paper, the process has received scant attention. The carbonates are not always correctly identified, and it is probable that ankerite is more common than is suspected. In the Coeur d'Alene district Anderson[4] believes that ankerite is the characteristic carbonate associated with the copper veins and that siderite is predominant only in the later lead veins. Siderite was formerly considered to be the only important carbonate in this district.

Chemical changes accompanying ankeritization are cited in three papers known to the writer. Junner's study[5] of the gold deposits of Victoria shows that carbonatization of diorite porphyrite involves a constant decrease in H_2O and an increase in K_2O and Fe_2O_3. At Kirkland Lake, Todd[6] describes carbonatization which, away from the fissures, involves an increase in CaO, MgO, FeO, and a decrease in K_2O, Na_2O, and SiO_2. Bruce and Hawley[7] present analyses of carbonatized greenstone at Red Lake which show marked changes in bulk composition, particularly of CaO, MgO, and Al_2O_3. In all these cases the carbonatized rocks have a direct connection with igneous intrusions, and changes in the quantity of the bases are not unexpected. In the Larder Lake area Cooke[8] states that, in addition to igneous affiliations, there is a structural control by which shear zones localize the process and provide channelways for the solutions.

ANKERITIZATION IN SOUTHERN QUEBEC

The chemical analyses of the metabasalt in Southern Quebec show that the formation of the uralite and ankerite phases involves no es-

[4] A. L. Anderson, "Sequence of Ore Deposition in Northern Idaho," *Econ. Geol.*, Vol. XXV (1930), p. 160.

[5] N. R. Junner, "Gold Occurrences of Victoria, Australia," *Econ. Geol.*, Vol. XVI (1921), p. 86.

[6] E. W. Todd, "Kirkland Lake Gold Area," *Ont. Dept. Mines*, Vol. XXXVII, Part II (1928), p. 91.

[7] E. L. Bruce and J. E. Hawley, "Geology of the Basin of Red Lake, District of Kenora," *Ont. Dept. Mines*, Vol. XXXVI, Part 3 (1927), p. 13.

[8] H. C. Cooke, "Kenogami, Round, and Larder Lake Areas, Timiskaming District, Ontario," *Can. Geol. Surv. Mem. 131* (1922), p. 52.

sential change other than addition of H_2O and CO_2. This has been repeatedly described for uralitized basalts, but is a new feature as far as the writer is aware in the case of ankeritization. These carbonate rocks have no connection with younger igneous intrusions and are not localized by shear zones in the metabasalt. In both chemical and field relations they differ from the ankerite rocks described from other localities, and their metamorphic history seems to be entirely dependent upon the alteration of the basalt.

The writer's conception of the post-magmatic changes in these rocks may be summarized as follows. The parent basaltic magma was the source of most of the H_2O and CO_2 which produced the mineralogic changes now apparent in the metabasalt. That part of it which was extruded contained, either initially or later in its history,[9] at least as much H_2O and CO_2 as now appears in the uralite rock analysis. The development of actinolite, epidote, and albite from the primary minerals indicates the part which H_2O played in this stage of metamorphism. At widely scattered centers in the uralite rock there accumulated unusual amounts of CO_2 which decomposed actinolite, epidote, and any remaining primary minerals, and fixed FeO, MgO, and CaO as ankerite.

These changes are endomorphic, i.e., they effect only the igneous body and do not alter the country rock. Field relations do not indicate any change in volume due to ankeritization, but if the assumption of exclusive CO_2 addition is correct it is not improbable that some volume increase has occurred. The temperature at which ankeritization occurred is probably lower than that necessary for uralitization but, on the other hand, is probably higher than that at which calcite would form in the uralite rock. Otherwise, calcite-chlorite aggregates would occur in place of the abundant ankerite, due to dissemination rather than fixation of FeO, MgO, and CaO. Calcite at the lower end of the temperature range and actinolite at the upper end of the temperature range of metamorphism are each mutually exclusive in their relations with ankerite. The ankerite rock contains neither of them and suggests an intermediate temperature of optimum development.

[9] Analogy with modern basalts, which are seldom uralitized, would favor a later introduction of H_2O and CO_2.

REVIEWS

La Géologie et les Mines de la France d'outre-Mer. Paris: Bureau d'Études Géologiques et Minières Coloniales. 1932. Pp. 604; figs. 38.

This volume on the geology of the French colonial empire corresponds in a general way to Reed's well-known *Geology of the British Empire*, which appeared in 1921. But, instead of being the compilation of a single author, the French work consists of contributions by many authors, each able to write authoritatively on a particular colony. The group includes a large number of names eminent in French geology.

The colonies are treated one after another in geographical order, the lengths of the chapters corresponding to the territorial and economic importance of the various colonies as well as to the extent of detailed geologic knowledge available. A geological map is given for each of the large possessions and geographical maps for the smaller ones. Geological maps in two colors cover Northern Africa (Morocco, Algeria, and Tunis) and Northern Indo-China on a scale of 1:3,000,000, and French Africa (Northern, Western, and Equatorial) on a 10,000,000th scale.

This volume, summarizing the present geological knowledge and mining developments of every French colony, is a most welcome and valuable addition to general geologic literature. Its value is enhanced by a short but selected bibliography at the end of each chapter. Most of these studies were delivered in a series of twenty lectures given by the authors themselves at the Museum d'Histoire Naturelle in Paris during the winter of 1931–32. The book is easy to read and the different portions are well co-ordinated in spite of the large number of contributors.

R. T. C.

The Tully Limestone of Central New York. By DAVID W. TRAINER, JR. *New York State Museum Bulletin 291*, Albany, 1932. Pp. 31; figs. 18, including 5 maps.

This brochure gives the results of a systematic microscopic study of a rock type which has a limited geographical extent, as the limestone becomes arenaceous eastward and changes westward into pyrite lenses. The

formation toward the west lies disconformably upon Hamilton shale and grades upward into Genesee shale.

The mineral composition is discussed in detail. Authigenic minerals are carbonates, chiefly calcite, pyrite, a clay mineral, and scarce barite, quartz, chlorite, and sphalerite (in Atrypa shells).

The varying amounts of carbonates were determined by carbon dioxide tests, and a "contour" map shows the distribution. Nineteen detrital minerals are reported in the insoluble residues, quartz, tourmaline, zircon, and muscovite being most common. Augite, plagioclase, hornblende, hypersthene, and monazite are found.

The quantity of insoluble residues is shown on a map by means of *isinsols*, here defined. By this means an old river mouth appears to be recognizable.

It is concluded that the formation was precipitated in a moderately warm, relatively shallow, foul sea of the marine littoral type. The detrital sediments were derived from igneous and metamorphic rocks of the Adirondack type.

This type of investigation should be much more common in order to interpret fully the sedimentary rocks, both as to their natural history and their uses. The author plans to study other New York formations in the same manner.

ARTHUR BEVAN

Records of the Geological Survey of India. Vol. LXV, Part I. Calcutta: Government of India, 1931. Pp. 24+187; figs. 9; pls. 2. 5*s*.

This volume contains the general report for the year 1930, and an "Additional Note on the Samelia Meteorites," by L. L. Fermor; two papers entitled "On the Zoning and Difference in Composition of Twinned Plagioclase Feldspars" and "The Albite Ala B. Twinning of Plagioclase Feldspars," by A. L. Coulson; and "Notes on Some Jurassic Fossils," by F. R. Cowper Reed.

After a discussion of personnel, equipment, publications, and the Empire Mining and Metallurgical Congress meeting the general report takes up paleontology, earthquakes, economic products, and the work of field parties. Thirty-four pages are devoted to a discussion of economic products in a number of localities, and seventy-six pages to other field work in 1930. In this section special stress is laid upon the petrology of the districts investigated. A valuable bibliography of the works devoted to Indian geology is included.

H. W. STRALEY, III